

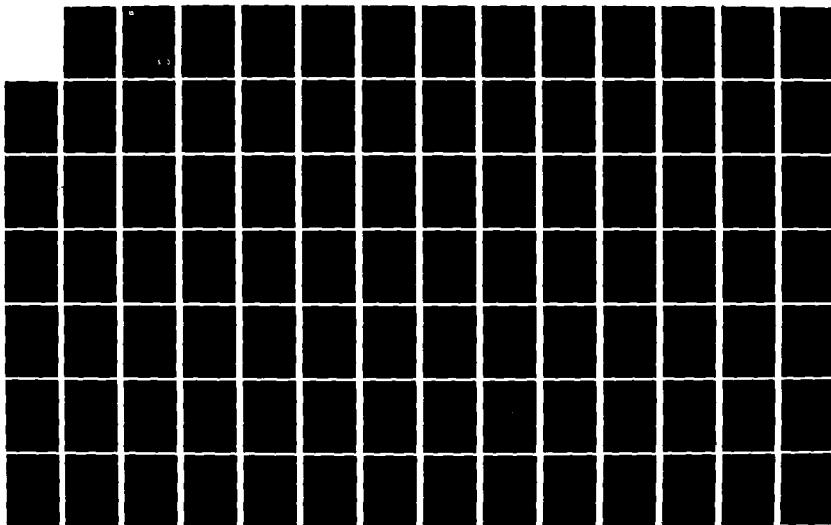
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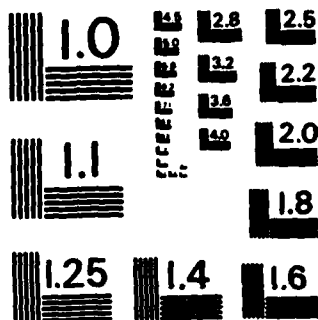
INVESTIGATION OF NEOTECTONIC ACTIVITY WITHIN THE LOWER
MISSISSIPPI VALLEY. (U) WATER ENGINEERING AND
TECHNOLOGY INC SHREVEPORT LA* S A SCHUM ET AL. SEP 82

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PHASE I
INVESTIGATION OF INDICTING ACTIVITY WITHIN
THE LOWER MISSISSIPPI VALLEY DIVISION

Polymology Program (P-4)

Report 2

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PREFACE

The study reported herein is a component of the Potamology Program (P-1) of the Lower Mississippi Valley Division. The Potamology Program is conducted under the direction of the Commander, Lower Mississippi Valley Division, and is a comprehensive study of physical forces which influence the flood carrying capacity and navigability of the lower Mississippi River. The purpose of the Potamology Program is to define cause-and-effect relationships that result in short-term and long-term changes in the lower Mississippi River's stage-discharge relationships and to develop improved design concepts and criteria for construction of channel stabilization works which will improve flood control and navigation along the lower Mississippi River.

The Potamology Program is composed of two major components: Sedimentation, Mississippi River Basin; and Aggradation and Degradation, Mississippi River. This study is Phase I of one item under the Aggradation and Degradation component. A future Phase II of this study will be directed toward investigations of specific reaches within the Vicksburg District including a structural process model to simulate uplift and/or subsidence.

The study reported herein was the responsibility of the U. S. Army Engineers, Vicksburg District, Vicksburg, Mississippi. Water Engineering Technology, Inc., of Shreveport, Louisiana, was contracted for the conduct of the study and Mississippi Experiment Station, Vicksburg, Mississippi, for the utilization of the report. The study was conducted during the period 1981-1982.

ACKNOWLEDGEMENT

Authors S. A. Schumm, C. C. Watson, and A. W. Burnett of Water Engineering Technology wish to convey appreciation and to acknowledge others who significantly contributed to the effort of this project. Messrs. James Tuttle, Brien Winkley, and Robert Rentschler, LMVD; and Mr. Don R. Williams, VXD, provided many valuable suggestions for the conduct and administration of the project. Mr. Roger Saucier and Mr. Lawson Smith, Waterways Experiment Station, provided thorough reviews of the draft report and helpful suggestions to enhance the final report. Dr. David Russ, U. S. Geological Survey, provided much of the New Madrid-Lake Co. Uplift material and a comprehensive review of the draft report. Mr. Sanford Holdahl and other staff members of the National Geodetic Survey provided instruction and advice in use of the survey data.

Dr. John Adams and Dr. Ray Frederking were involved early in the formulation of this study, and their assistance is acknowledged.

A National Science Foundation Project, No. EAR-7727573, provided partial support and valuable opportunity to examine the effects of neotectonic activity in other geographical areas.

Colonel Samuel P. Collins, Jr., was District Engineer during the preparation of this report.

SUMMARY

This report is the first in a series of a three-phase study of the effects of neotectonics upon the Mississippi River and tributaries. The objective of this preliminary study was to evaluate the hypothesis that gradual and presently continuing movements of the earth surface are affecting Mississippi River and tributary channel characteristics to such an extent that these effects should be included as design considerations for navigation and flood control projects.

A review of literature indicates that several major rivers are affected by tectonics. However, since alluvial channels are sensitive indicators of change in hydrology and sediment load and type as well as tectonics, the degree to which rivers are controlled by tectonics alone has not previously been thoroughly investigated. The examples discussed in the literature of tectonic control of river behavior deal primarily with rather dramatic movements, earthquakes, and with the influence of rocks of different resistance emplaced along the channel. The literature has been less specific about the effects of gradual and continuing crustal movements in a large alluvial river such as the Mississippi.

This report provides evidence to indicate that the effects of crustal movement in a river system can be categorized as follows:

- 1.) Change in watershed drainage pattern
- 2.) Channel aggradation or degradation
- 3.) Change in channel pattern or sinuosity
- 4.) Channel diversion or avulsion
- 5.) Flooding due to subsidence.

This preliminary report demonstrates that at least three major geologic uplift features continue to be active in the Mississippi Valley and that influence of these features may impact present navigation and flood control features.

Geologic and precise level surveys indicate that crustal uplift of about 3 millimeters per year can be expected at some locations within the Mississippi Valley. The average low water reference plane gradient of the Mississippi River is only about 90 millimeters per mile, or about half the width of this page per mile. It is easy to understand the significance of a gradual 3mm/year movement that is accumulated over a project life of 50 years. Further, it can be seen that the same accumulation of uplift could be a significant cause for peculiarities of channel behavior which develop over a period of years in a channel reach that otherwise has been free of problems.

This is the report for the preliminary phase of this study. Further phases of the study are to be directed to additional definition of the effects of crustal movement, to a specific investigation of particular locations, and to development of design criteria encompassing these long term effects in planning for navigation and flood control projects.

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CHAPTER 1

INTRODUCTION

Neotectonics is defined as "the study of the last structures and structural history of the earth's crust" during the later Tertiary and Quaternary (American Geological Institute, 1972, page 477). The recognition of tectonic events and deformation of the earth's surface, during the last few million years of earth history is, of course, an important function of the geologist and geomorphologist, but the objective of neotectonic studies is to identify those areas where there is deformation by uplift, subsidence, or faulting. Neotectonics has affected the earth's surface as we know it, but little consideration has been given to the subtler aspects of neotectonics, especially as it relates to rivers, river behavior and river morphology. Geomorphologists have studied drainage network patterns and anomalous reaches of rivers that are affected by geologic structure, but the effect of neotectonics on alluvial rivers has received little attention. For example, Ollier (1981, p.180) states that "the degree to which rivers control and are controlled by tectonics remains to be worked out and the elucidation of the interplay of geomorphology and tectonics will require a degree of geomorphic input that is not yet available."

The objectives of this study are: 1) to review the international literature and to determine the effect of neotectonics on alluvial rivers, 2) to deduce the effects of neotectonics of various types on alluvial rivers, 3) to determine if neotectonic activity is influencing

the Mississippi River between Cairo, Illinois and the Gulf of Mexico.

In order to provide the background necessary for a study of the possible effects of neotectonics on the lower Mississippi River a review of the neotectonic literature will be presented in Chapter 2. In Chapter 3 the existing relationships between river morphology and the effect of valley-floor gradient on channel patterns will be used to deduce the types of changes that can be expected as alluvial rivers adjust to deformation. In the final portion of this report evidence will be presented for deformation of the alluvial valley of the Mississippi River and on the effect of this deformation on the Mississippi River and some of its tributaries.

Structural Landforms

There are two types of structural landforms (Twidale, 1971): 1) Primary or tectonic landforms, which are due directly and only to activities within the earth's crust, without the intervention of the forces of erosion. These primary landforms are clearly very young and erosional activity has not significantly modified their morphologic characteristics; 2) Secondary landforms, which are due to the modification of the tectonic landforms by erosional forces. Distinction between the primary and secondary structural landforms is clearly one of age. Eventually all tectonic landforms will become secondary landforms that have been modified by erosional and depositional processes.

Structural landforms can be readily recognized in consolidated rocks, where stream channels and drainage networks have incised into and have adjusted themselves to the varying resistance of rocks which compose the earth's surface. The best examples are the various types of

drainage networks, for example, a rectangular drainage network forms as the result of reorganizing joint sets or faults, and a dendritic drainage network develops on bedrock strata (Fig. 1-2, Table 1-1).

The geomorphologist is interested with long-term adjustment of drainage patterns to structural influences. In such cases he views the effect of structure and topography as an equilibrium field. Nevertheless, there has been some long periods of adjustment, as the channels in the drainage networks react to structural influences. If deformation was too rapid, undoubtedly there was a disruption of the existing river system. If deformation was slow the existing river system could persist in its location, but the changes of valley-floor slope could require an adjustment of river gradient.

It is not surprising that little attention has been given to the effects of neotectonics on rivers because variations of channel characteristics can usually be attributed to downstream variations in discharge, sediment load, and the type of sediment moved through the channel (Schumm, 1977) or to local geology (Hess+G, 1967, p. 2746). Nevertheless, drainage anomalies that may be due to neotectonics have been identified by many investigators. These drainage anomalies consist of local channel pattern change, local widening or narrowing of channels, anomalous ponds, meanders or glacial fills, decreased channel gradient, variation of natural levee width and discontinuous levees. Finally, any anomalous curve or turn is suspect. An active dome may force a channel to adopt a course around it, and a fault may offset the channel laterally (Fig. 1-2).

Many of the major rivers of the world follow structural lines and major geofracture systems (Potter, 1971). On a continental scale many

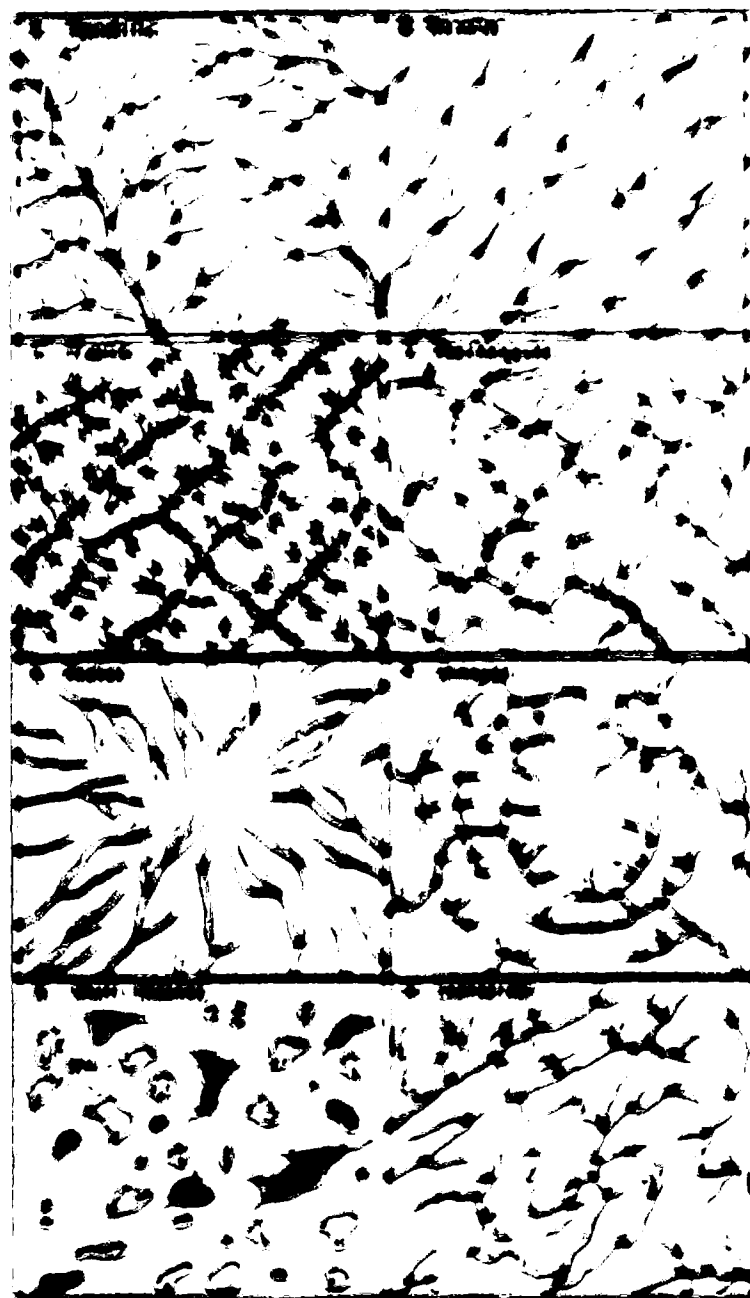


Figure 1.1. Tissue Preparation Procedures. See Table 1.1
(From Figure 1.1.1)

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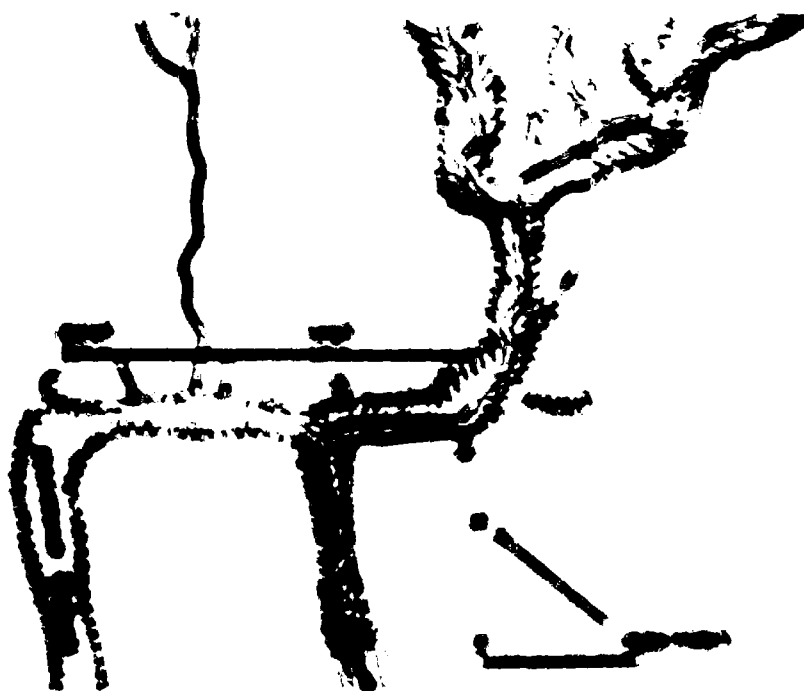


Figure 1

Sketch and working drawing of engine pump for the test
 made from drawings 1977

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THESE STATES ARE SHOWN IN THE FOLLOWING PATTERNS:
 WESTERN STATES - GRID
 MOUNTAIN STATES - CROSS-HATCH
 SOUTHERN STATES - DIAGONAL LINE
 EASTERN STATES - DOTTED PATTERN

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Figure 1

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Figure 6

In addition to all of these structural features, the entire valley may be tilted upstream or downstream or the tilting may be across the valley toward either side of the floodplain. The possibilities are great, but in reality the result will be local steepening or reduction of gradient or cross-valley tilting.

Summary

Sufficient work has been done to demonstrate that neotectonic deformation of alluvial valleys occurs at rates that will affect alluvial stages. The deformation will cause a reduction or increase of valley floor gradient with corresponding changes of channel gradient.

CHAPTER 2

REVIEWS AND EXAMPLES OF NEOTECTONIC EFFECTS ON RIVERS

The clearest evidence for neotectonic effects on rivers will be anomalous reaches showing dramatic changes of pattern and gradient that cannot be attributed to other causes. Examples of such river changes have been reported and in order to provide a background from which the study of the Mississippi River can be considered, a review of the world-wide literature is presented here. It should be noted that the interest of earth scientists in neotectonics is increasing rapidly, as is the activity of the U. S. Geological Survey regarding earthquake studies and neotectonics (Hadley and Devine, 1974 and Howard, et al., 1978). In addition, the International Association for Quaternary Research (INQUA) has recently established a commission on neotectonics. In fact, the first symposium on recent crustal movements was held in Moscow in 1936 and the sixth international symposium was held at Stanford University in 1977 (Whitten, et al., 1977). In 1967, Hiersmann (1967) published a 28 page bibliography dealing with recent crustal movements (neotectonics) in Europe.

General Studies

Three books have been published in the last decade that deal with structural landforms (Tricart, 1974; Twidale, 1971; Ollier, 1981). They are particularly weak, however, in the discussion of neotectonic effects on rivers.

Tricart's (1974) book was first published in French in 1968. It includes a short section on "active faults and the hydrographic system". The discussion centers on long-term effects of faulting and warping, the

offsetting of river courses and the formation of lakes by faulting. He stresses the effect of faulting on meanders, which may become very angular in plan. He also cites the example of the Hutt River in New Zealand (Tricart, 1974, p. 227) which incised from two to three meters and ceased to be navigable, as a result of the 1855 earthquake and the tilting that resulted.

Twidale (1971, p. 133-136) recognizes the effect of faulting on drainage lines. He states that the rise of a fault block across a stream causes either impedance of drainage and the formation of a lake or swamp, or the diversion of the stream and the development of an irregular or abnormal drainage pattern. The displacement along faults is irregular and unequal, and sag ponds due to marked local subsidence are a common feature of fault zones. Many of these features are described from the San Andreas fault zone in California (Sharp, 1954). Twidale refers to the Murray River in the Echucha district of Victoria, Australia as a classic example of tectonic diversion caused by the rise of the Cadell Fault block (Fig. 2-1). This impressive example of channel pattern modification by tectonic activity is described by Bowler and Harford (1966). An uplifted fault block in the Riverine Plain near Echuca has converted the Murray River from a single channel stream to an anastomosing system of channels that flow around the obstruction. The abandoned segment of Murray River is preserved on the dipslope of the fault block.

Ollier (1981) devotes a chapter to drainage patterns, rivers and tectonics, and he discusses the effects of warping and faulting on drainage systems (Fig. 2-2; 2-3). Figure 2-2 illustrates some of the results of warping in an area of dendritic drainage patterns. At 1 an

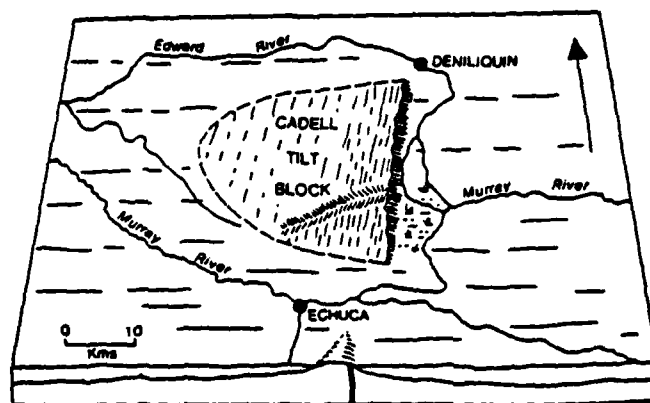


Figure 2.1 Disruption of Murray River by Cadell Fault
(from Ollier, 1981)

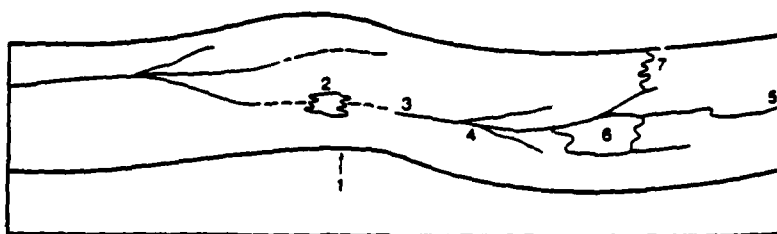


Figure 2.2 Effects of Warping On Drainage
(from Ollier, 1981)

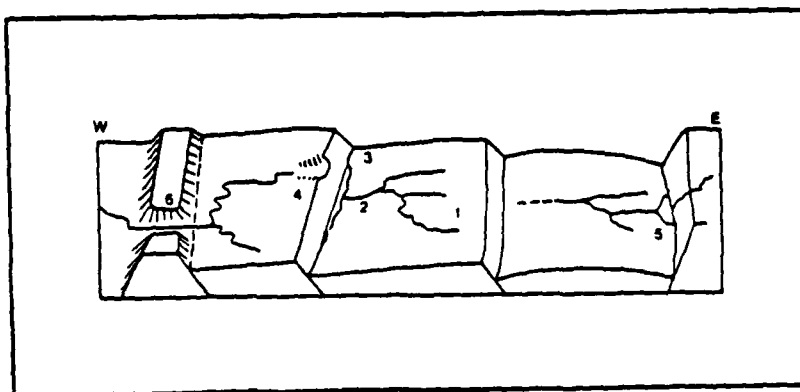


Figure 2.3 Effects of Faulting on Drainage
(from Ollier, 1981)

upwarp has disrupted the drainage. Downstream from the bulge, the river flows with a reduced discharge. At 2 on the flat crest of the bulge, if it is broad and gradient is low, swamps or shallow lakes may develop. The drainage is reversed at 3 and flows away from the uplift. Tributaries still flow into the main channel but they are oriented upstream and form a barbed drainage pattern at 4. The original drainage direction is maintained at 5 and drainage into a lake at which 6 formed on the upstream side of the uplift. Flow from the lake follows the depression at 7 and parallels the uplift. If the stream had been able to maintain itself across the uplift, it would be an antecedent stream, as shown at 6 on Figure 2-3.

Figure 2-3 shows the effects of faults (Fig. 1-4) on a stream with drainage disruption at 1 and 2. A lake has formed at the base of the fault scarp at 2, and the drainage is diverted along the fault at 3. The faults have beheaded the river at 1 and 4, but the fault at 5 has only steepened the headwaters. The disruptions of drainage networks, as shown in Figures 2-1, 2-2, and 2-3, are extreme examples of the effects of deformation, and they are easily recognized as such.

During geologic time, the effects of faulting and folding can be substantial. Freund, et al., 1968, have evidence that movement along the Dead Sea Rift in Israel has disrupted streams draining from Jordan. These streams crossed what is now the Dead Sea Valley to the Mediterranean Sea. Portions of the channels of these rivers are now displaced about 43km laterally due to movement along the boundary faults of the Dead Sea Rift Valley.

Changes of an entire drainage network, as a result of tilting, are described by Sparling (1967) who notes that isostatic adjustment in

Ottawa County, Ohio has increased the gradient of some streams and decreased the gradient of others. Incision of the steepened streams permitted them to capture the streams that were aggrading as a result of reduced gradient.

Doornkamp and Temple (1966) describe the formation of lakes as a result of gradient reduction on the eastern side of the rift valley near Lake Victoria. Other streams in the area have been steepened and their longitudinal profiles show a series of nick points as a result of that rejuvenation.

Russell, (1936) describing the various drainage patterns found in the flat alluvial lands of Louisiana, recognizes a network pattern of poorly developed drainage that is then converted to a dendritic pattern as a result of tilting and steepening of the gradient (Fig. 2-4).

Valleys and Terraces

Deformation will affect valleys and terraces as well as drainage patterns. The most convincing evidence of deformation is the tilting of alluvial terraces in a valley. If the deformation persists, the oldest terrace is the most deformed by uplift or subsidence, and it will show the greatest offset by faulting.

VALLEY DEFORMATION

Deformation of the bedrock floor of a valley will be indicated by bedrock configuration and the thickness of overlying alluvium. For example, Kowalski and Radziokowska (1968) note that alluvium will be thickest over down faulted blocks (graben) and thinnest over areas of uplift, as expected.

In areas of subsidence, streams may have broad valleys with well developed flood plains and meandering channels (Sizkov and Zfumster,

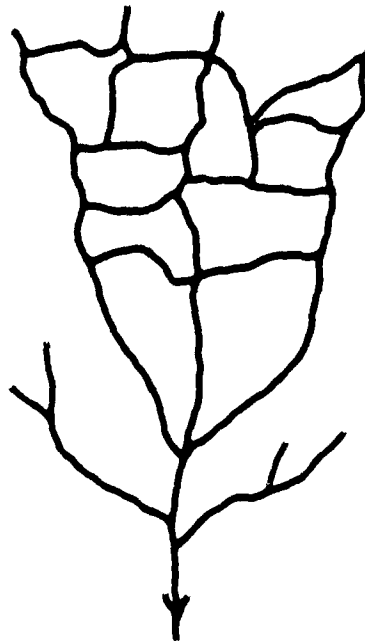


Figure 2.4 Conversion of poorly defined drainage network
into a dendritic pattern by slope increase.

1967). This is presumably in contrast to younger valleys that, as a result of river incision, are narrower and more likely to be braided.

STRESS ADJUSTMENT

The presence of the valley itself can lead to further deformation and further enlargement of the valley. For example, river incision into massive bedrock may lead to the formation of large scale exfoliation joints. These pressure-release joints parallel the canyon walls in the Colorado Plateaus of western United States (Bradley, 1963) and they cause enlargement of the canyons by rockfalls and weathering. Bradley (1963) reports that in massive sandstones, joints that parallel the canyon wall are most frequent within 7m of the wall, but beyond 15m there are no joints parallel to the canyon wall. Obviously, these fractures are due to the excavation of the canyon itself and to the relief of stress in the wall rocks.

At dam sites in the Ohio Valley, Ferguson (1974, 1981) reports compressive faults in the valley bottoms and tension fractures in valley walls. The tension fractures are nearly vertical and they parallel the valley walls in a manner similar to those observed in the Colorado Plateaus. The bedrock valley bottoms have been subject to compressive forces produced by fracturing and horizontal expansion of the valley walls. This horizontal force causes failure in the form of arching, thrust faulting and bedding plane faults in the valley floor.

In the Allegheny Plateaus of eastern United States, the stress release and natural bottom-fracturing and daming have been encountered in almost all projects that exposed the sides and bottoms of river valleys. In addition, a differential stress, as a result of unloading that is generated by canyon erosion, has produced valley anticlines in

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The opposite case occurs where crustal depression has occurred as a result of surface loading by water or ice. For example, Lander (1985)

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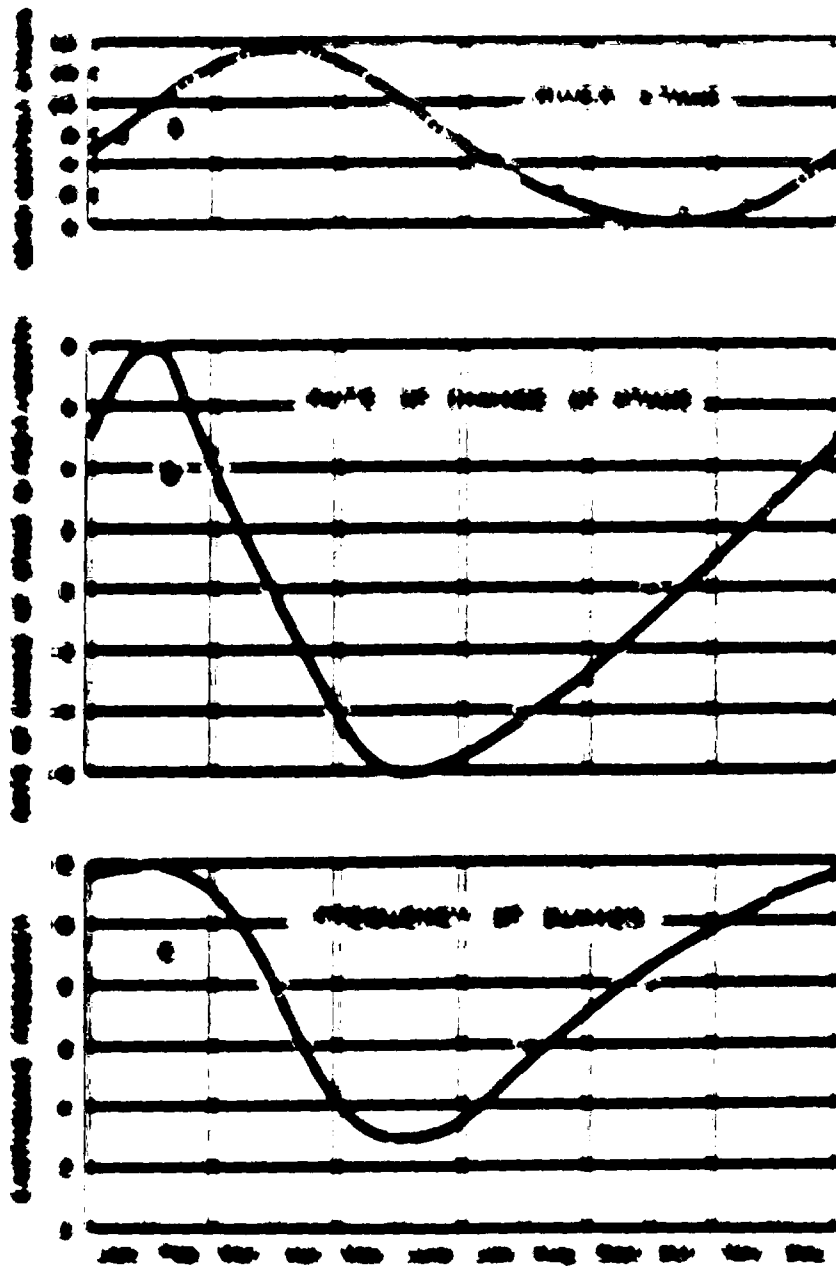
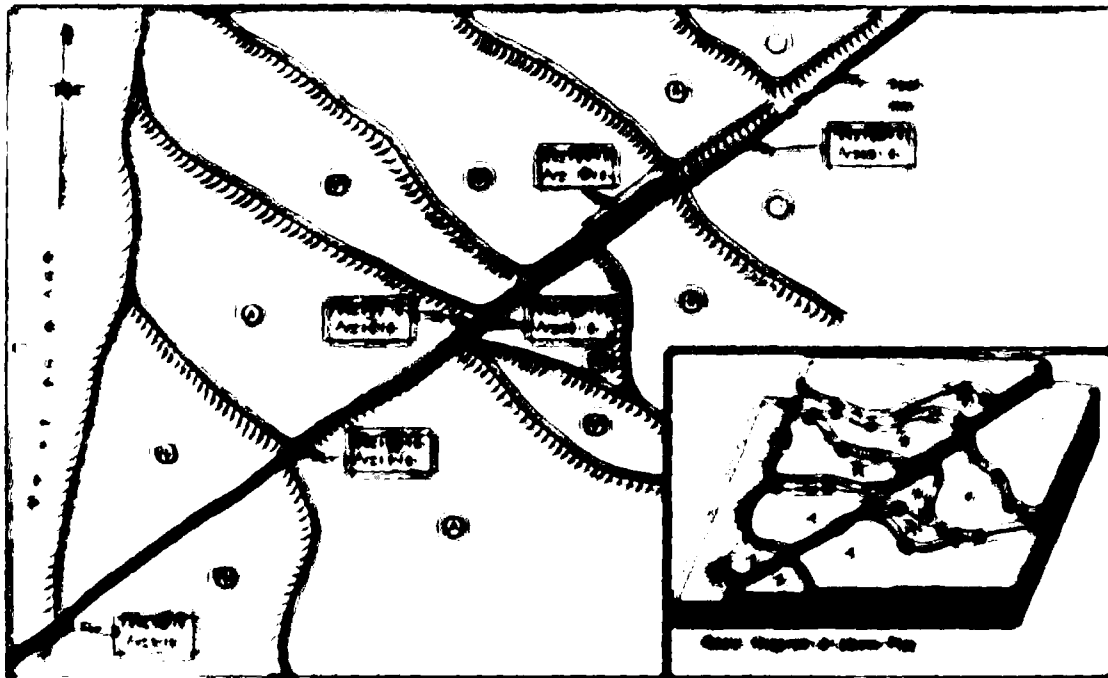


FIGURE 1. (a) Relative intensity of sound as a function of frequency. (b) Relative intensity of sound as a function of frequency. (c) Relative intensity of sound as a function of frequency. (From McGraw-Hill, 1953)



The map shows a network of roads and paths. The main road is labeled "Main Ave" at the top center. There are several other roads branching off from this main road. The roads are labeled "Main Ave" on the left side, "Main Ave" at the bottom center, and "Main Ave" on the right side. There are also labels for "Main Ave" and "Main Ave" on the right side. The map shows a complex intersection of roads with various paths branching off. An inset map in the bottom right corner shows a larger area with a grid of roads and a central intersection. The inset map is labeled "Main Ave" and "Main Ave".

can be reflected only in those geomorphic features that react to the smallest changes of slope, the gradient of terraces and streams, and meander characteristics. In addition, the variations of thickness and distribution of recent sediments indicate the variability of neotectonics in a given valley. Clearly, one of the most sensitive indicators of change is the valley floor profile and longitudinal profile of the stream (Bondy, et al., 1967; Zuchowicz, 1979.)

A means of detecting deformation of a valley floor is by use of the specific stage concept (Volkov, et al., 1967). That is, as uplift occurs, both the river bottom and the water level at a specific discharge will fall as the channel deepens. This can be detected at a gauging station when long records are available. There are, of course, other reasons for channel incision, and it will be rare that changes in stage height can be related to neotectonics alone.

Stream patterns should be very sensitive indicators of valley slope change and this will be considered in detail in the next chapter, but Adams (1980) has demonstrated a relation between measured tilt rates and downstream changes of sinuosity of the type described by Schumm (1972, 1977). That is, in order to maintain a constant gradient a river that is being steepened by a downstream tilt will increase its sinuosity whereas, a reduction of valley slope will lead to a reduction of sinuosity or degradation and aggradation if the pattern cannot change.

Interesting examples are provided by the Mississippi River between St. Louis and Cairo and the lower Missouri River (Adams, 1980) as well as the Red River of the North near Winnipeg, Canada (Vanicek and Nagy, 1980).

Tufdale (1966) reports that both the Flinders and Leichardt Rivers

have changed to a braided pattern as a result of the steepening of their gradient by the Selwyn Upwarp in northern Queensland, Australia.

Veronenko and Ivanov (1977) have considered the effect of tectonics on meanders in more detail than most investigators. They reviewed the Russian literature and conclude that rivers crossing uplifts are deeply entrenched and have a minimum number of meanders of small amplitude. The largest number of meanders occur upstream of the structures.

Another aspect of meander growth and river shift is the lateral or transverse tilting of the valley. In general, a river should shift in a down tilt or a down slope direction and concentrate its attack on the valley side that has been down tilted (Cotton, 1941). However, a more recent study by Hanson (1980) shows that the south-flowing Beatton River in British Columbia is affected by isostatic tilt to the east. This has caused a deviation from the normal downstream migration of the meanders such that there is an average 20° eastward or leftward departure from the flow direction. The tilting has augmented the easterly-directed flow velocities resulting in an easterly bias to channel migration. However, in spite of this progressive eastward migration, the present channel lies closer to the west side of the valley. Although the meander loops tend to migrate toward the east, frequent channel cutoffs isolate oxbows on the east side of the flood plain and leave the present channel flowing closer to the west wall of the valley. Meanders of the Beatton River are confined within the valley walls and, therefore, the river is not free to shift laterally. This is unlike the situation described by Mike (1975) for the Hungarian plain where the Tisa River has shifted laterally for long distances during recent geologic time.

As meanders shift down valley, their movement may be retarded if

more resistant alluvium or bedrock is encountered. Hence, a fault may present a barrier with the result that upstream meanders are compressed and deformed (Fig. 3-6). This has been documented by experiment (Gardner, 1973, 1975). A similar pattern will result if the river encounters bedrock as it crosses an upwarp or if as a result of incision it encounters resistant materials in a portion of its course.

Examples of River Response

Two areas of the world that are occupied by big rivers have clear evidence of neotectonic activity. These are the Tigris and Euphrates Valley in Iraq and the Indus Valley of Pakistan.

Ambraseys (1978) has been involved in seismo-tectonic studies throughout the Middle East. He writes that, although major earthquakes have been documented in this region and their effects have been described, there are probably many small earthquakes that are associated with surface deformation that have never been documented. This is particularly relevant to the southern Mississippi Valley where, because of the lack of seismograph stations, the record is inadequate.

Ambraseys (1978) believes that much surface deformation is aseismic in the Tigris and Euphrates Valley. In places, surface and underground canals have excavated into their old beds as a result of uplift. Elsewhere, erosion could not keep pace with the rise of the canal bed and the water supply system was abandoned, today showing a reverse gradient. Underground canals have been sheared off by faulting. Some of them have been repaired, but, today these new alignments are now offset by a few meters. For example, North of the Persian Gulf in the southwestern corner of Iran is the Shaurn anticline. This forms a range of low hills that emerge from the alluvial plains between Shush and Ahwaz. Folding

began in late Pliocene and it still continues. In the first or second century AD two canals were cut across the anticline in order to lead water from a canal system on the northeast flank to the more extensive and fertile plains on the southwest. These channels afford a unique opportunity for measuring the uplift of the anticlines since the canals were built seventeen hundred years ago. One canal still carries water, but, where it crosses the anticline it has cut down about 4 meters below its original bed. The more northwesterly canal crosses a higher part of the anticline and the axial section had to be tunneled for a short distance. In this case, either the flow of water was not strong enough to enable the canal to excavate its bed or else its maintenance was neglected and it was abandoned. An accurate survey along its alignment shows that along the anticlinal axis the bed of the canal has risen at an average rate of approximately 1 meter per century (Lees, 1955).

Lees (1955) recognizes subsidence at the head of the Persian Gulf. He identified an old canal system that presumably irrigated an extensive area that is now submerged beneath the waters of the Persian Gulf. In contrast to the usual explanation for the shift of the shoreline as a result of deposition, Lees concludes that the Tigris and Euphrates Rivers are not constructing a normal delta, rather they are discharging sediment load into a tectonic basin. Subsidence is episodic, and in the intervals the depressions fill up with sediment. In this same area Mirjayar (1966) states that the left bank terraces of the Euphrates River are substantially lower than the right bank terraces, apparently due to movement of the valley floor.

Another example of tectonic effects is in Iraq where the character of the ancient irrigation systems associated with the Diyala River, a

tributary of Tigris near Baghdad, has been altered by upwarping. Uplift during the last thousand years caused incision of the Diyala River into its alluvial plain and abandonment of irrigation canal systems (Adams, 1965).

There are numerous active faults in the Indus Valley (Kazmi, 1979). The most spectacular effect of active faulting is in the Rann of Cutch fault zone. In this region of the lower Indus Valley in 1819 a severe earthquake resulted in the uplift of a ten mile wide and fifty mile long tract of alluvial land with a relief of about twenty feet. This feature was locally known as Alah Bund (Oldham, 1926). The eastern branch of the Indus was blocked by the formation of the Alah Bund (Oldham, 1926). The channel at that time was dry, but flow was reestablished during a flood in 1828. If the river had been flowing it is possible that the channel would have maintained itself.

Lyell (1857, p. 462) states that "for several years after the convulsion of 1819, the course of the Indus was very unsettled, and at length, in 1826, the river threw a vast body of water into its eastern arm, forcing its way in a more direct course to the sea, burst through all the artificial barriers which had been thrown across its channel, and at length cut right through the Alah Bund". For discussion of recent history of the Indus see Holmes (1968) and for an interesting hypothesis concerning the decline of an Indus civilization see Dales (1966) who suggests that Indus Valley cities were flooded as a result of major tectonic activity in the Indus Valley.

Precise Leveling Survey Data

A source of information from which evidence of neotectonic activity can be ascertained is the wealth of data compiled by the National

Geodetic Survey (NGS) and the National Ocean Survey (formerly U. S. Coast and Geodetic Survey). This type of information is perhaps more familiar to the engineer, and for the areas where comparative surveys have been made for an adequately long period of time, these surveys provide data of geological significance.

The primary use of the precise leveling survey data is to determine the vertical movement of a bench mark over a period of years, thus a velocity in millimeters per year (mm/yr.) can be computed. Along a survey line, the relative velocity of two bench marks determines a tilt rate for the land surface along the survey route. The tilt rate is expressed in radians per year (RAD/YR).

Some degree of subjectivity is inherent in the use of the data. The data furnished by the NGS consist of plotted and tabulated values of relative elevation change between successive surveys along the same survey route. It is the pattern of movement rather than absolute elevation or absolute velocity which is of primary importance (Brown & Oliver, 1976). Thus it is a subjective decision as to the length along the survey profile that represents a significant pattern.

In most areas there are not sufficiently multiple relevelings to allow for the development of a clear picture of the short term temporal behavior of vertical crustal movements. In addition, where multiple relevelings are available, the apparent temporal behavior is often quite complex. In many cases it is not presently known whether this complex behavior is due to actual crustal movements or to survey errors.

LEVELING DATA ERROR

Brinker and Taylor (1965) categorize leveling errors into three sources: instrument, natural, and personal. Instrument errors pertain

to those errors which may occur due to inaccurate equipment, i.e., inability of level instrument to sight along a horizontal plane or incorrect level rod length. Careful calibration of the equipment can minimize these effects.

Personal errors in instrument readings, improper plumbing of the level rods, or other faulty operation of the equipment can be controlled by adherence to proper operating procedures. Routine closure looping procedures minimize the likelihood of misreadings in the field.

Natural error sources include refraction characteristic changes due to sudden atmospheric changes. Errors due to refraction tend to be random over a long period of time, but could be systematic for a single day (Brinkler & Taylor, 1965). Temperature variation causing heat waves and expansion and contraction of equipment, and wind induced equipment vibration can be significant, but operational procedures can minimize these error sources. Thurm (1971) examined the effects of refraction, tidal influence, and thermal rod change over a level net and concluded that correction for these factors did not significantly affect elevation change measurements.

Brown and Oliver (1976) point out that the most serious type of errors are systematic errors which tend to accumulate with distance along the survey route. An example of this type of error is a small error in level rod length which accumulates with every foresight. In utilizing the NGS tabulations and plots of elevation change between comparative surveys, this type of systematic error will indicate a steadily increasing bench mark velocity as distance from the survey line initiation point decreased. Data exhibiting this tendency should be immediately suspect.

Random errors are propagated as the square root of the number of measurements made (Brown & Oliver, 1976). Recognizing this and assuming a constant length of measurements, the NGS has set criteria to define the accuracy of leveling work as a function of the square root of the distance of the leveling distance, expressed in kilometers.

Leveling observations in tectonically active areas, while in some cases poorly understood, are often accepted as geologically significant. Leveling differences in stable plate interiors have been interpreted as indicating vertical motion, although tectonic activity is generally unexpected.

Some fundamental questions have been raised concerning the reliability of leveling estimates of vertical movement. In some areas the trends of geodetically measured movements are consistent with trends in the geologic record, although the contemporary rates of movement are ten to one hundred times faster than average rates estimated from geomorphic and geologic evidence for the past one to ten million years. This apparent contradiction is termed the rate paradox. The paradox has led to the hypothesis that contemporary movements are episodic or oscillatory with relatively short periods on the order of a hundred thousand years. Episodic movements are those associated with earthquakes. Oscillatory movements may result from magnetic activity, that is, inflation and deflation. Adams and Reilinger (1980) conclude that although there are high tilt rates and relatively rapid uplifts in mid-continent U. S. these observations appear to represent real crustal movements. The level lines follow large rivers, and there is a correlation between tilt and downstream changes in river sinuosity (Adams, 1979). The mid-continent is apparently being deformed by

oscillatory tilting with a period of about three thousand years. Even without the above interpretation there is good evidence that present tilting has continued in the same direction for at least eighty years and if the periodicity of three thousand years is even approximately correct, the assumption of a constant rate of tilting for a hundred years seems justified (Adams and Reilinger, 1980).

Another example of confirmed uplift is in the Rio Grande Valley north of Socorro, New Mexico where leveling lines show doming above an active magma body (Reilinger and Oliver, 1976). This is associated with high micro-earthquake activity and high heat flow. The observed uplift rates can be explained by inflation of the magma body where uplift at the center of the dome is more than 4mm per year, relative to the side. In the same area, Pliocene sands of the ancestral Rio Grande River are tilted to the south. If the doming occurred at the presently measured rate then the sands would have had the same downstream slope as the Rio Grande 50,000 years ago. There is a general consistency between assumed magma flow rates and the volume of the magma chamber and the total tilt shown by the Pliocene sands.

The releveing studies indicate that in areas free of major earthquakes, the movement rates that have been obtained from releveing observations spanning tens of years, although fast in the geologic sense, can likely be extrapolated for fifty years with some degree of assurance (Adams and Reilinger, 1980).

Summary:

The literature review indicates that neotectonic activity is worldwide. As Potter suggests, big-river valleys are zones of crustal weakness, where river morphology and behavior are affected by neotectonic

activity.

A variety of evidence exists for neotectonic activity ranging from precise repeat-survey data to the deformation of terraces and man-made features and river pattern anomalies.

CHAPTER 3

EFFECT OF SEDIMENT MOVEMENTS ON ALLUVIAL CHANNELS

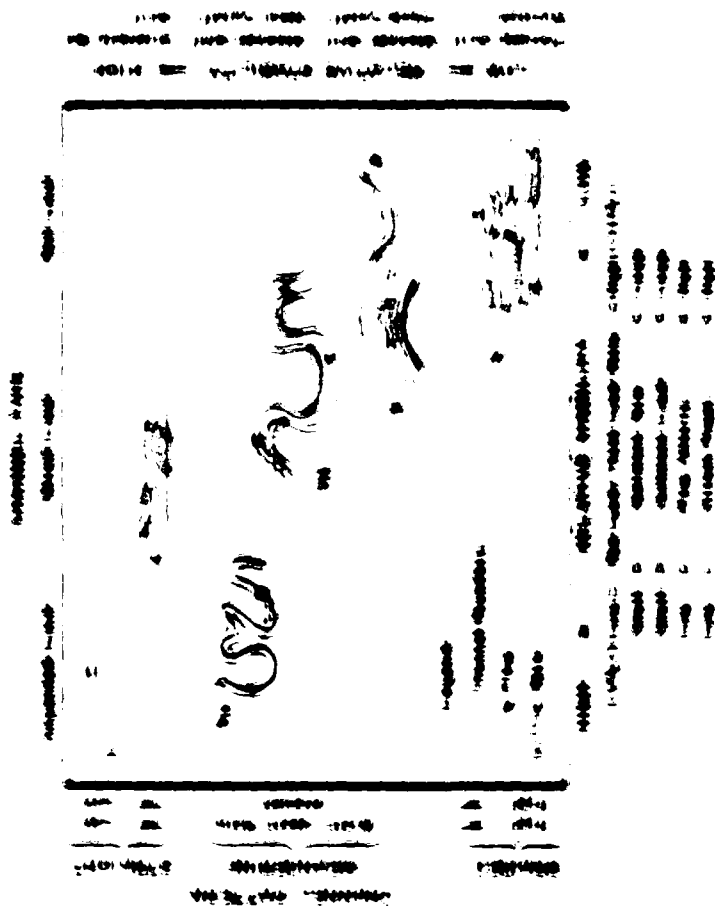
The preceding discussion and review showed that sediment activity can significantly control river patterns and behavior and that this is most pronounced when rates of shifting resistance are balanced by feeding (Fig. 3-1). The literature is less specific about the effect of recent channel movements on alluvial rivers with as the threshold therefore, it is necessary to review how an alluvial river would respond to the types of deformation outlined in Chapter 1 (Fig. 3-2).

If the surface of an alluvial valley is supplied by major fault movement the results will depend on that movement. A major lateral movement will offset the channel (Fig. 3-2). Feeding that produces a break in the profile of a stream will cause either incision or deposition depending on the nature of the movement (Figs. 3-2, 3-3, 3-4).

Progressive movements may have little immediate effect, but the cumulative effect of changing the shape of the alluvial valley can be great, and the exposure of sediments of shifting resistance in the bed of the channel should cause a change of channel morphology.

It is apparent that different types of alluvial channels will respond differently to deformation; therefore, the characteristics of alluvial channels must be reviewed before their response can be evaluated. This can best be done by discussing a simple classification of alluvial channels that is based on type of sediment load and pattern (Fig. 3-5).

Five basic channel patterns exist (Fig. 3-5): 1) straight channels with either aggrading sand bars, 2) or aggrading alternate bars with



OFFICE OF THE SECRETARY OF THE ARMY
WASHINGTON, D. C.
JANUARY 1, 1917
SIR:
I have the honor to acknowledge the receipt of your letter of the 29th ultimo, in relation to the proposed construction of a new road from the existing road to the new road, and in reply to inform you that the same has been referred to the proper authorities for their consideration.
Very respectfully,
Your obedient servant,
J. H. HARRIS, Secretary.

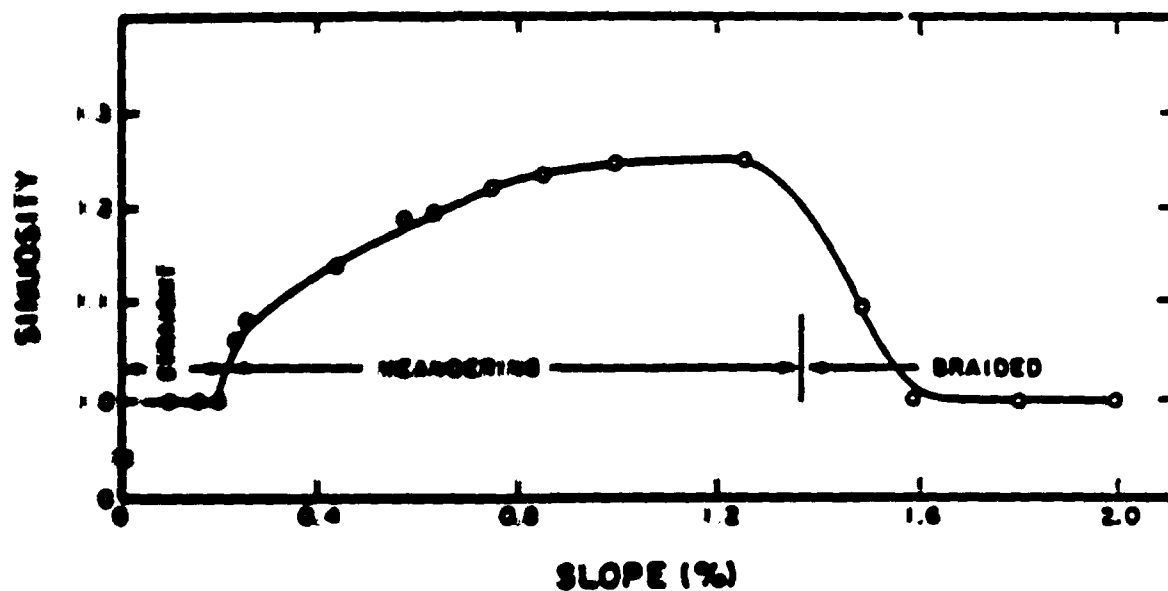
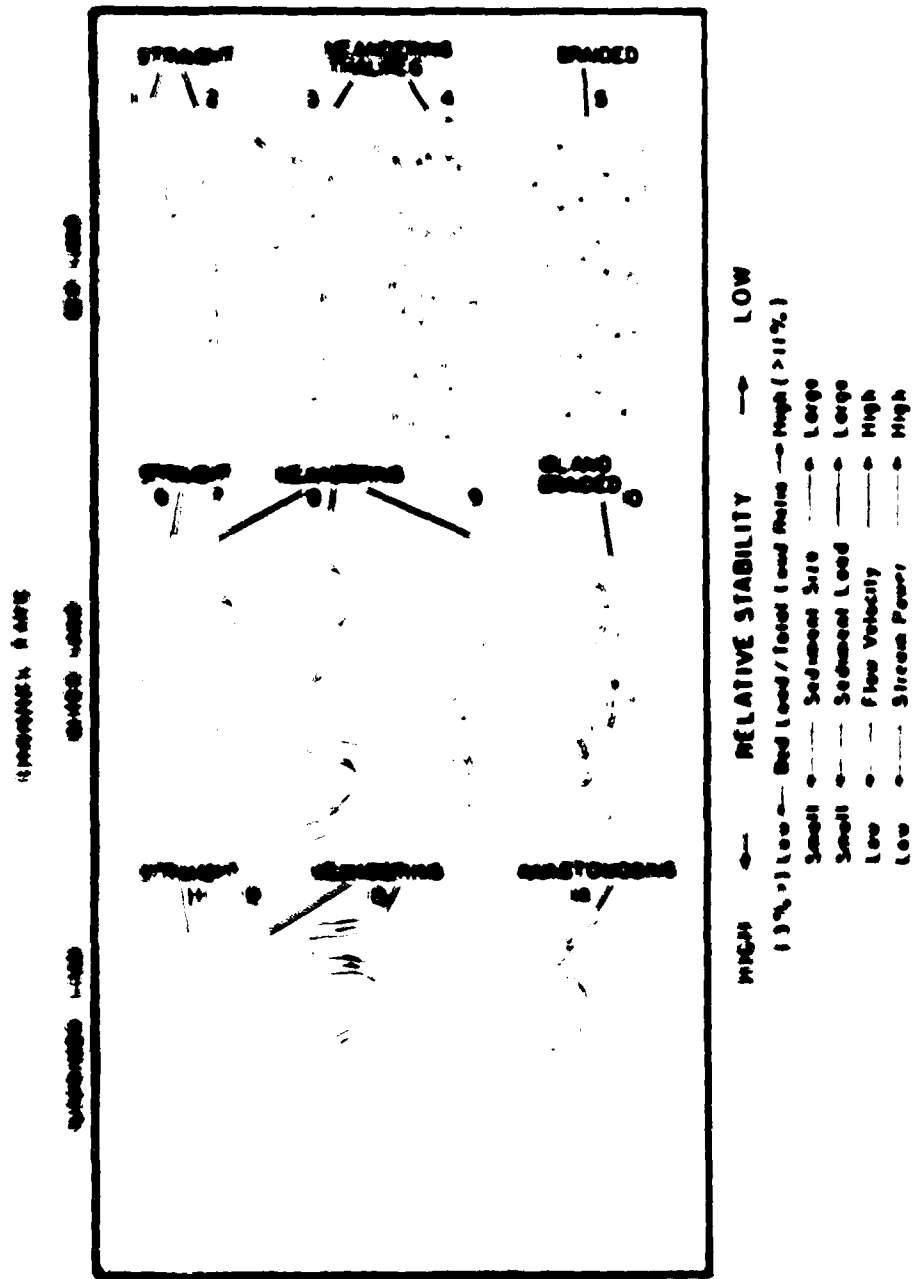


Figure 3.2 Relation between Flume slope and sinuosity during experiments at constant water discharge. Sediment load, stream power, velocity increase with flume slope and a similar relation can be developed with these variables (from Schumm and Khan, 1972).

CHANNEL PATTERNS



Channel patterns are determined by the relative stability of the channel bed and banks. The relative stability is a function of the bed load, total load, sediment size, sediment load, flow velocity, and stream power. The diagram shows that channel patterns change as these factors vary. For example, a straight channel becomes meandering as the sediment load increases, and a meandering channel becomes braided as the flow velocity increases.

As compared to the bed-load channels, the five-mixed load channels (Fig. 3-3) are relatively narrower and deeper, and there is greater bank stability. The higher degree of bank stability permits the maintenance of narrow, deep, straight channels (Pattern 6), and alternate bars stabilize because of the finer sediments to form slightly sinuous channels (Pattern 7). Pattern 8 is a truly meandering channel, wide on the bends, relatively narrow at the crossings, and subject to chute cut-offs. Pattern 9 maintains the sinuosity of a meandering channel, but with a larger sediment transport the presence of bars gives it a composite sinuous-braided appearance. Pattern 10 is a braided channel that is relatively more stable than that of bedload channel 5, and in fact, it may be a transitional pattern between the bar-braided pattern 5 and a narrow, more sinuous channel (Pattern 8). Under present conditions it is an island-braided channel.

Suspended-load channels (Fig. 3-3) are narrow and deeper than mixed-load channels. Suspended load Pattern 11 is a straight, narrow, deep channel. With only small quantities of bed load, this type of channel may have the highest sinuosity of all (Patterns 12 and 13). The steepest suspended-load channel (Pattern 14) may be anastomosing. Bars will not form because bed-load transport is so low, but multiple channels will develop to produce the anastomosing pattern that is characteristic of some fine-sediment alluvial plains. The anastomosed channels may be kilometers in length. Therefore, the intervening vegetated areas are not simply islands of the Pattern-10 type. On a smaller scale, a very similar appearing pattern develops when, as a result of reduced flood peaks and annual discharge there is a metamorphosis from bed-load Patterns 4 and 5 to a single channel pattern

(e.g. Patterns 7 and 8). Such changes have been documented for the Platte and Arkansas Rivers. (Nadler and Schumm, 1981).

The most common of the above patterns are bed-load Patterns 4 and 5, mixed load Patterns 7 and 8, and suspended-load Patterns 11, 12, and 13. Some patterns are not common because they are readily converted to other patterns. For example, the stabilizing effect of vegetation, especially hardy pioneer species such as willows and cottonwoods, will convert Patterns 2 and 3 to Patterns 7, 8 and 9. The closure of side channels will convert Pattern 14 to Patterns 12 and 13.

Rivers may undergo a metamorphosis during which the channel morphology changes completely; that is, a suspended-load channel (Pattern 12) could become braided (Pattern 5), or a braided channel (Pattern 5) could become meandering (Pattern 8 or 12), etc., when there is a sufficiently great change in the type of sediment load transported through that channel. Therefore, the change from one type of channel pattern to another may be relatively common, as the nature of the sediment moved through the system changes and this may be simply the effect of tributary sediment contributions (Schumm, 1977), or upstream degradation or aggradation.

Changes of valley-floor gradient provide another explanation of downstream pattern variations. Variations in the valley floor slope such as those in the Mississippi River Valley have influenced sinuosity, as the river adjusts its pattern to maintain a constant gradient over the changing valley-floor slope (Schumm, et al., 1972).

Variations of valley floor slope can be the result of several influences. Tectonic activity may change the slope of the valley floor and have its effect on the channel pattern (Adams, 1980). In addition,

a high-sediment-transporting tributary may build a fan-like deposit in the valley, which will persist even after the tributary sediment load has decreased. When the main river crosses this fan, pattern changes will result, as the river attempts to maintain a constant gradient. Tributaries to the Jordan River have developed fan-like deposits in the valley, and the valley floor of the Jordan Valley undulates as a result. The Jordan River, as it approaches one of these convexities, straightens as it crosses the upstream flatter part of the fan and then it develops a more sinuous course on the steeper downstream side of the fan (Schumm, 1977, p. 140).

It is important to realize that channels that lie near a pattern threshold (Fig. 3-2) may change their characteristics dramatically with only a slight change in the controlling variable. For example, some rivers that are meandering and that are near pattern thresholds become braided with only a small addition of bed load (Schumm, 1979).

Experimental studies and field observations confirm that a change of valley floor slope will cause a change of channel morphology. The change will differ, however, depending where the channel lies on a plot such as that of Figure 3-2 and depending on the type of channel (Fig. 3-3).

In order to simplify the discussion of neotectonic effects on alluvial rivers, only changes of valley floor slope and stream gradient that are associated with doming will be considered. The decreased gradient upstream of the axis of a dome is similar to the reduced slope downstream of the axis of a basin (Fig. 1-4i) and upstream of a reverse fault (Fig. 1-4c). The increased gradient downstream of the axis of a dome is similar to the increased slope upstream of the axis of a basin

(Fig. 1-4i) and downstream of a normal fault (Fig. 1-4b).

Listed on table 3-1 are pattern changes to be expected for each of the fourteen patterns of Figure 3-3, as valley and channel slope is reduced and then steepened by doming.

The reduced slope may either cause a pattern change as indicated or aggradation. Aggradation may cause shallowing of the channel and avulsion or simply development of a braided pattern. The increased slope will cause a pattern change, as indicated (Table 3-1), or if the channel cannot adjust in that manner, degradation will result. If aggradation or degradation results, the reach of channel affected will be obvious with aggradation in the flattened reach upstream and below the reach of degradation. Degradation will occur in the steepened reach and it will be extended upstream with time (Fig. 3-4).

Figure 3-5 shows the pattern change expected for initial Patterns 4, 7, 9 and 12 (Fig. 3-3), which are typical patterns for each of the three channel types (Fig. 3-1).

As noted above, such pattern change may have other causes, but such pattern anomalies are indications of river reaches that are affected by neotectonics.

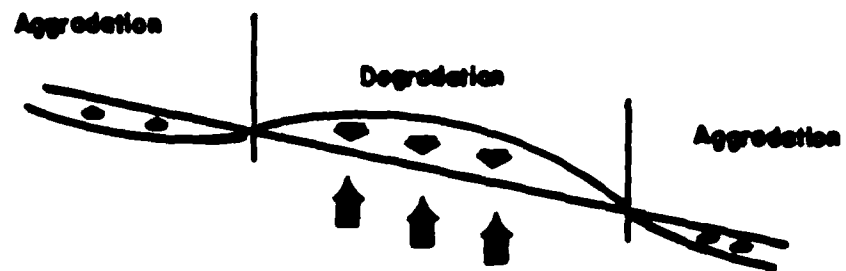
Although pattern changes may dominate river response, deposition in reaches of reduced gradient is likely as is channel incision and bank erosion in reaches of steepened gradient (Fig. 3-4). In fact, upstream deposition will reduce downstream sediment loads thereby increasing the tendency for downstream erosion.

Channel erosion at reaches of steepened gradient may produce an excess of sediment that is delivered downstream to a reach that has not been affected by uplift (Fig. 3-4). In this case, the increased

Case	Dome		(active)		
	Channel normal	Gradient normal		reduced	increased
1	A		2 or D	1 or A	
2	1 or A		3 or D	2 or A	
3	2 or A		4 or D	3 or A	
4	3 or A		5 or D	4 or A	
5	4 or A		5 or D	5 or A	
6	A		7 or D	6 or A	
7	6 or A		8 or D	7 or A	
8	7 or A		9 or D	8 or A	
9	8 or A		10 or D	9 or A	
10	9 or A		10 or D	10 or A	
11	A		12 or D	11 or A	
12	11 or A		13 or D	12 or A	
13	12 or A		14 or D	13 or A	
14	13 or A		14 or D	14 or A	

Table 3.1 Effect of uplift on 14 river patterns. A indicates aggradation. D indicates degradation. The numbers refer to the patterns of Fig. 3.3

A Dome



B Fault

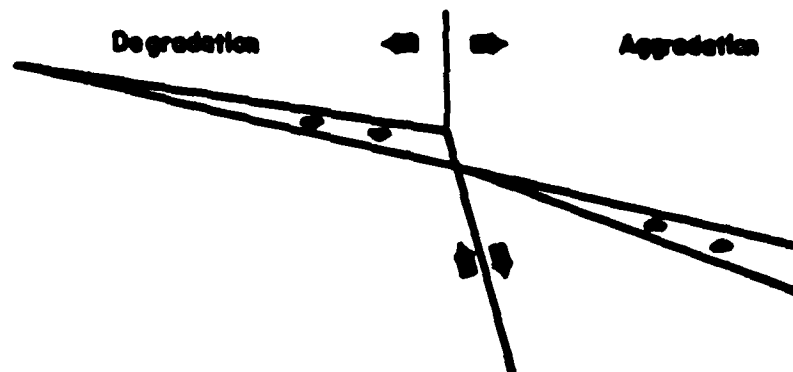


Figure 3.4 Reaches of degradation and aggradation associated with uplift.

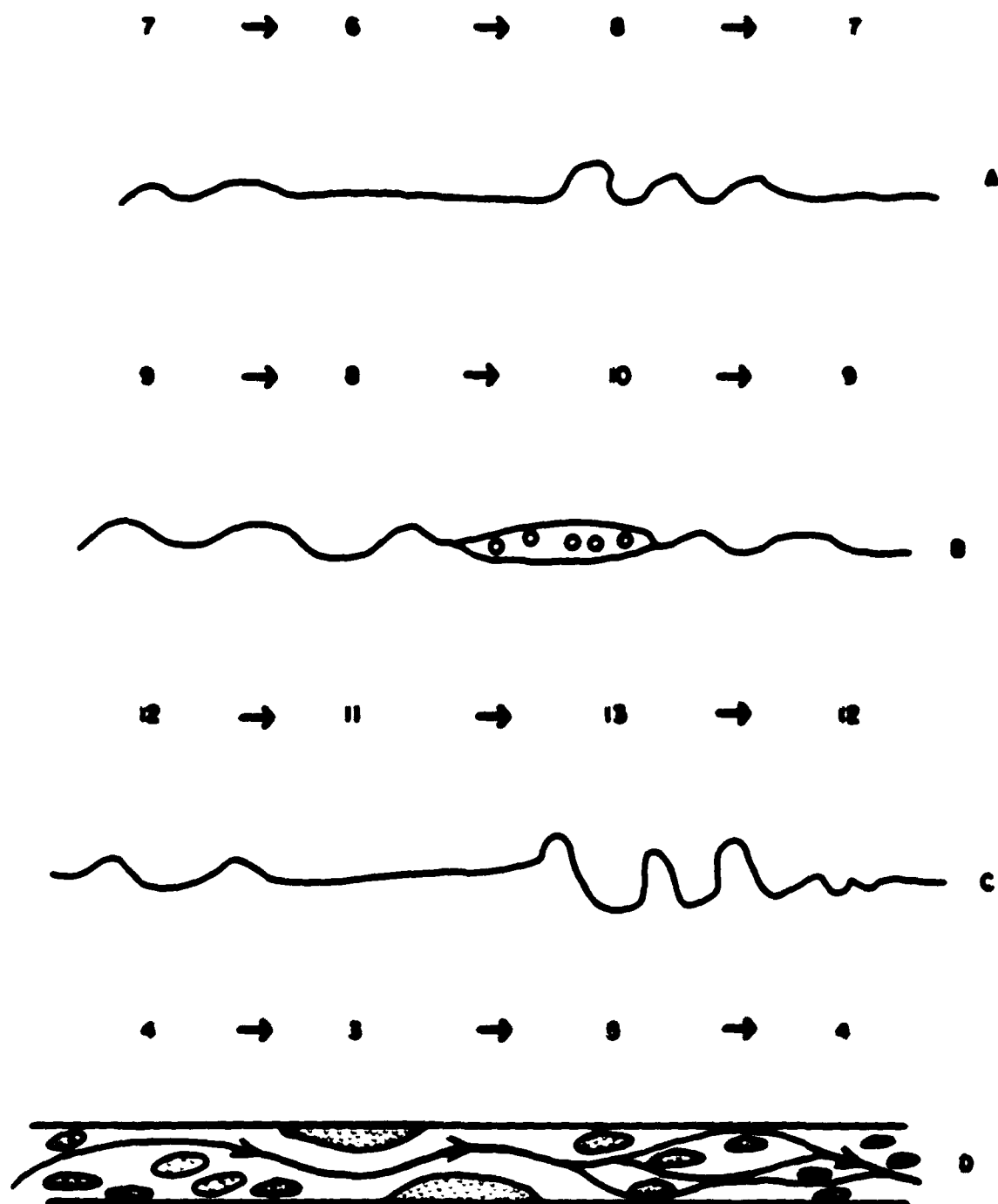


Figure 3.5 Examples of pattern change based upon changes indicated from Table 3.1. Numbers indicate river patterns. (Fig. 3.3)

sediment load may induce aggradation, meander cutoffs and even metamorphosis from meandering to braided patterns. Therefore a local neotectonic effect may lead to further channel adjustments both up and downstream. Therefore, it is possible that after an episode of deformation the greatest activity will be upstream (incision) and downstream (aggradation) of the site of deformation.

Another aspect of neotectonics is the emplacement of either wetter or more resistant materials in the channel. Figure 3-6 is an attempt to illustrate the effect of such sediment changes on the channel. More resistant sediment will confine the channel and retard meander shift and bank erosion. The result should be deformed or compressed meanders upstream of the resistant material (Fig. 3-6a) and a change of meander character at the contact.

Introduction of less resistant sediments should lead to bank erosion, channel widening and perhaps a local conversion to a mixed-load or bed-load channel type of morphology.

If an alluvial valley is sufficiently wide, as in the Mississippi Valley, the river may not be maintained over the deformation, especially where unroofing is occurring the river may shift laterally off the structure. As described earlier (Chapter 1) such a river will have an anomalous bend in its course as it passes around the structure (Fig. 3-7a). In addition, lateral tilt in a wide valley should, in spite of Hansen's (1975) observations, shift a channel to the down-tilt side of the valley (Fig. 3-7b) (Adams, 1980). In addition, on an alluvial plain deformation may lead to avulsion and shift of meander belts as the course of a river is obstructed by local uplift (Fig. 3-7a2).

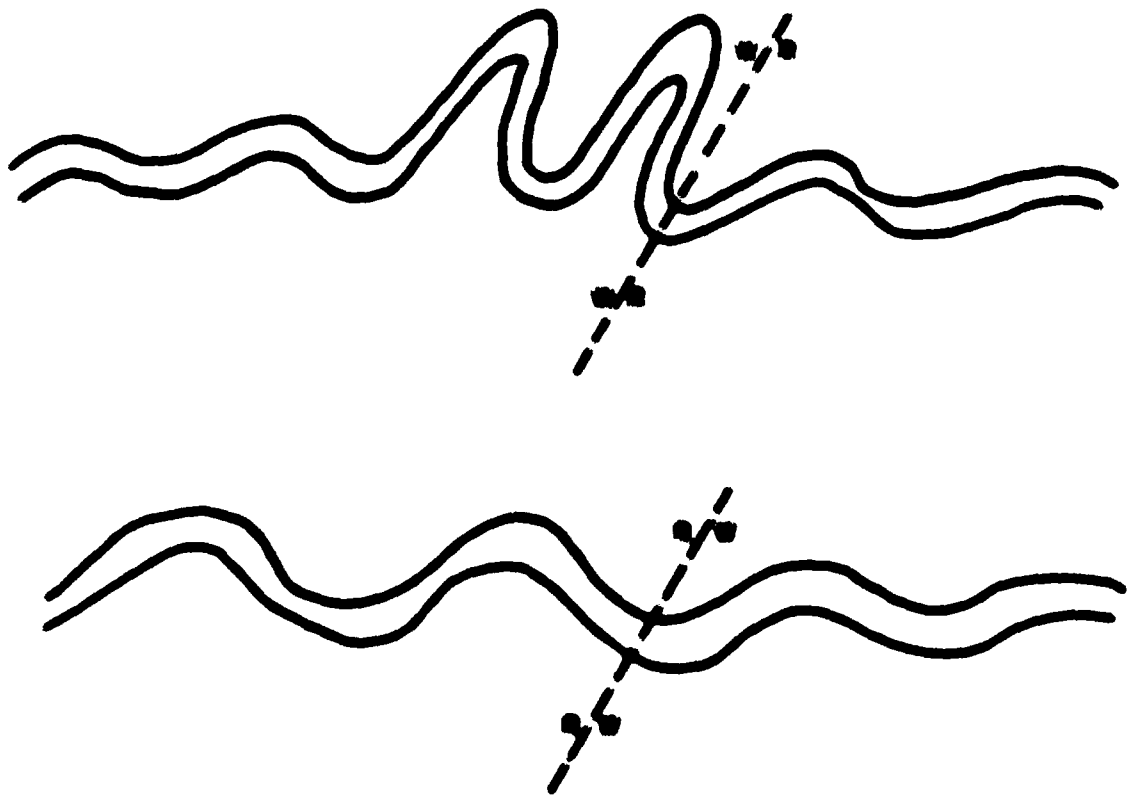


Figure 20 Effect of change of material on wave speed
 W - Wave v - Wave Speed

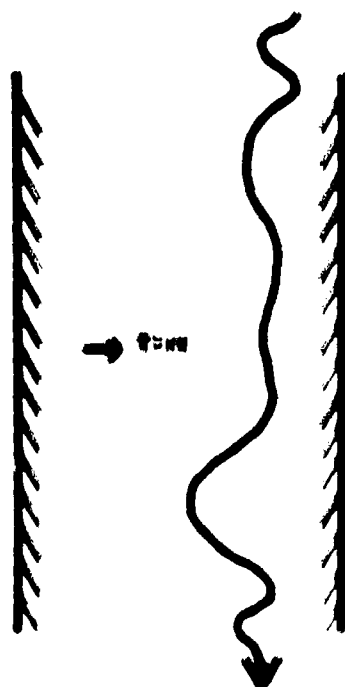
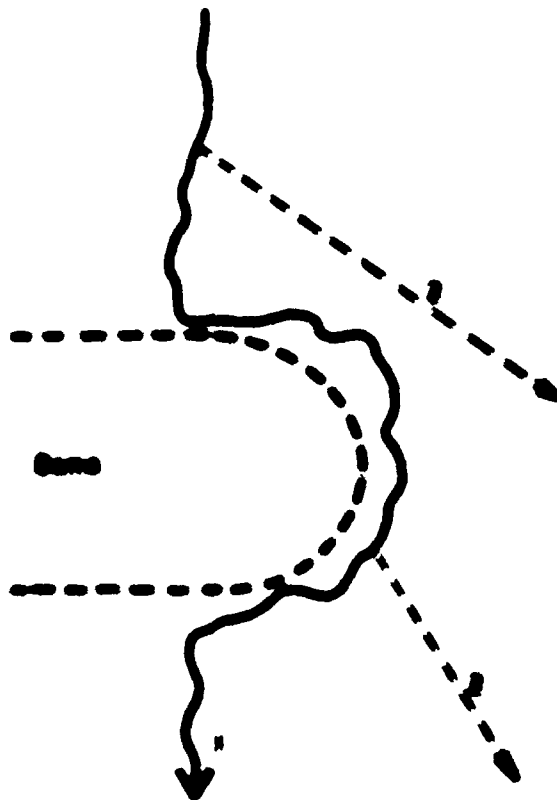


Figure 37

Effect of depth (A) and velocity (B) on stream position

CHAPTER 3

DESCRIPTION OF THE MISSISSIPPI BASIN

The following is a summary of a description of Mississippi drainage and related areas given by Monroe (1902). This section provides a framework for the more detailed discussions of various basins in the upper, middle and lower Mississippi.

The Mississippi drainage includes the central portion of the northern half-circle of latitude 30° north to 40° north, on either side of the Gulf of Mexico. It is bounded on the north by 40° north, on the east by the Gulf of Mexico, on the south by 30° north, and on the west by the Rocky Mountains. It is bounded on the north by 40° north, on the east by the Gulf of Mexico, on the south by 30° north, and on the west by the Rocky Mountains. It is bounded on the north by 40° north, on the east by the Gulf of Mexico, on the south by 30° north, and on the west by the Rocky Mountains. It is bounded on the north by 40° north, on the east by the Gulf of Mexico, on the south by 30° north, and on the west by the Rocky Mountains.

The Mississippi drainage is a broad, somewhat symmetrical system that extends from the north to the south. It is bounded on the north by 40° north, on the east by the Gulf of Mexico, on the south by 30° north, and on the west by the Rocky Mountains. It is bounded on the north by 40° north, on the east by the Gulf of Mexico, on the south by 30° north, and on the west by the Rocky Mountains. It is bounded on the north by 40° north, on the east by the Gulf of Mexico, on the south by 30° north, and on the west by the Rocky Mountains.

The axis of the drainage runs southeast from the Rocky Basin, and then turns south and southeast around the eastern side of the Monroe basin where the system is constricted between it and the Jackson Dome in Mississippi. The system is symmetrical in this area especially to

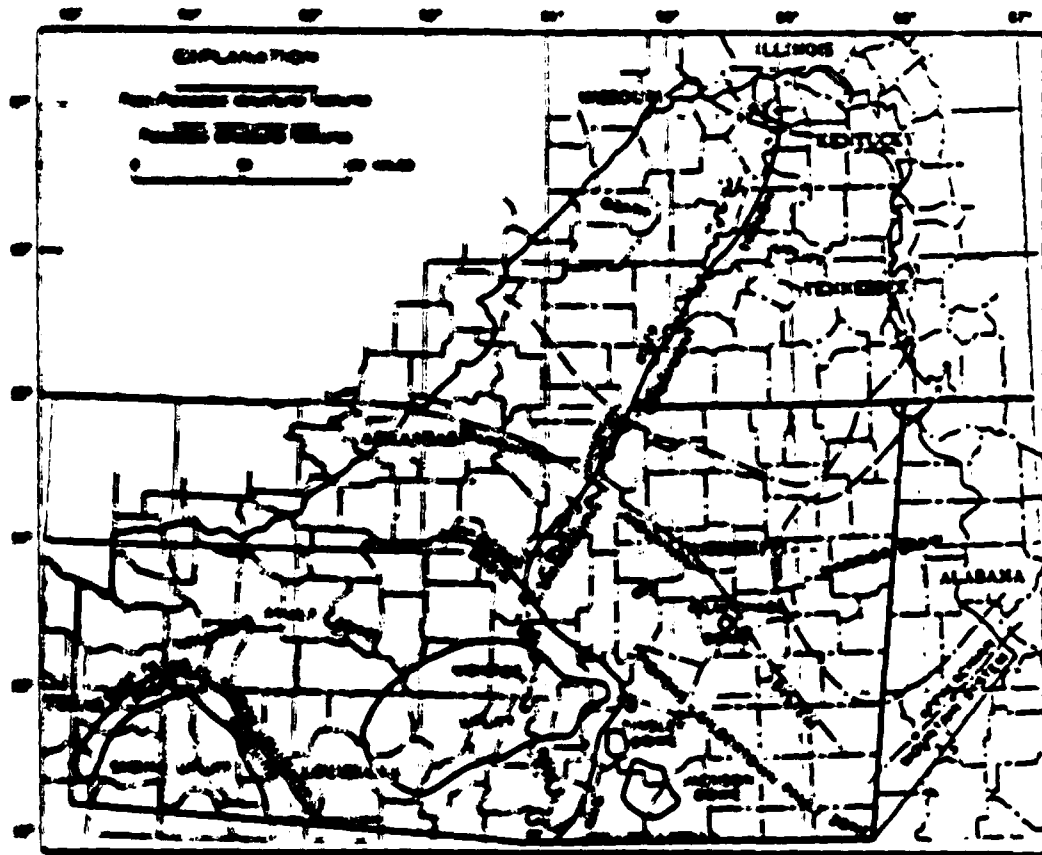


Figure 5.1 Structure map of the Mississippi embayment. (from Cushing, et.al., 1964)

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(The following information was obtained from the records of the Federal Bureau of Investigation.)

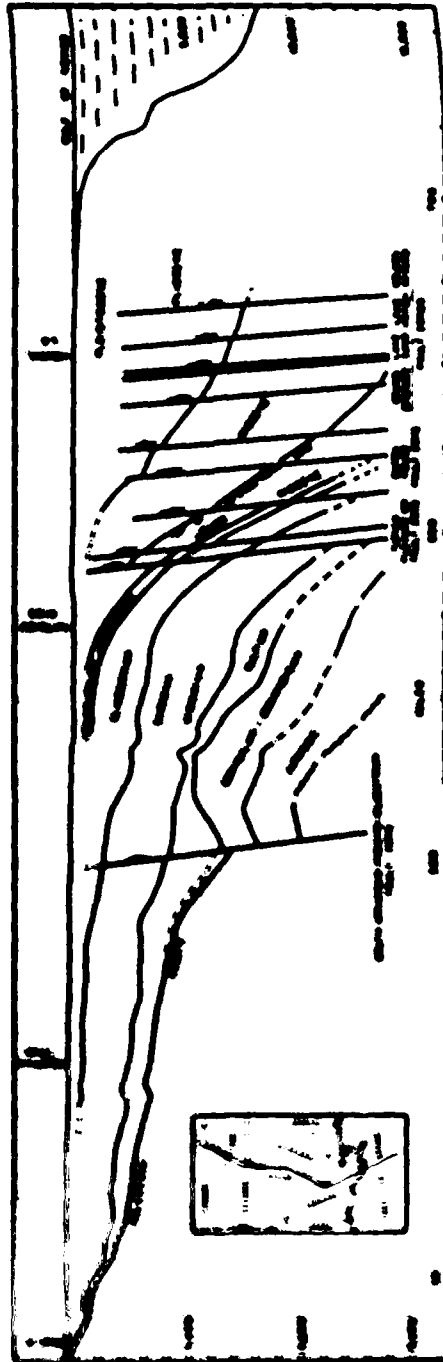


FIG. 4.2. Segments of the San Andreas fault system showing possible oil fields. (Modified after Rik, 1944.)

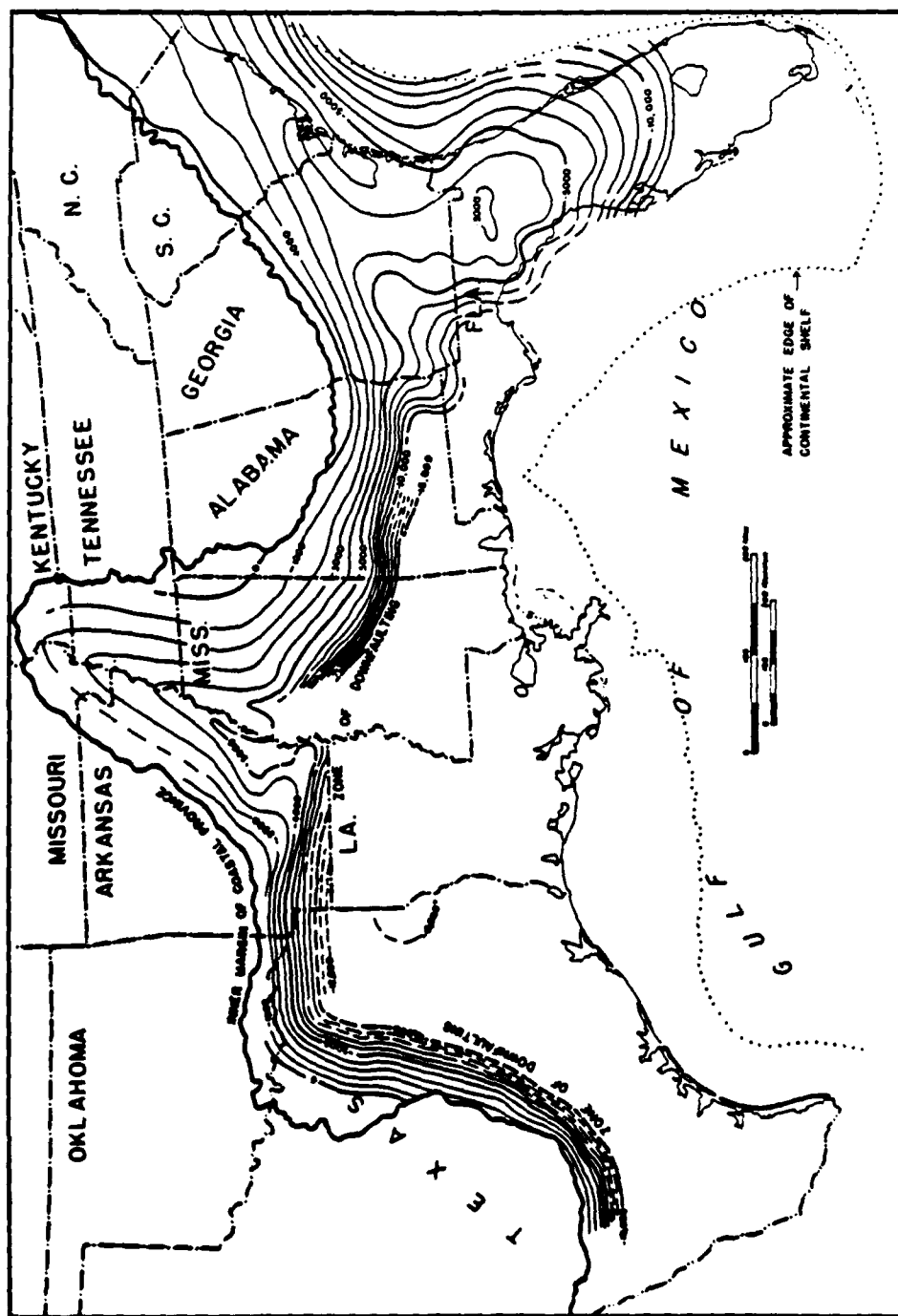


Figure 4.3 Generalized structure on top of pre-coastal basement rocks. (In part after Woollard, Bonini, and Meyer, 1957.)

southeast and northeast, southwest orientations (Fig. 4-4). However, Saucier (1979, unpublished report) states that intensive geologic studies along several fault zones have failed to substantiate their presence and that there is little evidence that faulting has been a significant factor in determining Mississippi River position and morphology. The exception is in the deltaic plain where east-west trending faults have been mapped. They are probably growth faults associated with downwarping of the Gulf Coast geosyncline (Fig. 4-5).

The Monroe uplift, a major structure feature, is in the middle valley. The modern uplift is a broad, relatively flat-topped dome. In the lower part of the valley to the east is the Wiggins uplift. The term is here applied to the irregular partly-arcuate, bifurcating positive element in southwestern Alabama, southeastern Mississippi and eastern Louisiana. The limits of this uplift on the west and south are unknown and it is intriguing to speculate that it may extend below the river to the west. The feature arcs to the west in the vicinity of the Mississippi state line, maintaining a westerly trend into Stone County, where it appears to bifurcate. One lobe extends to the northwest, the other, the southern lobe commonly called Hancock Ridge, is thought by some geologists to be a continuation of Appalachian structural trends. Gulfward in Mississippi Sound, Lake Borgne, Chandeleur Sound and adjacent areas of Louisiana is a gentle southwesterly trending arch that appears to be a continuation of the southwesterly trend of Hancock Ridge. It plunges to the southwest and disappears beneath the St. Bernard Parish, Louisiana.

The Mississippi embayment appears to be structurally similar to the East African rift valleys and other rift valleys in that it has undergone

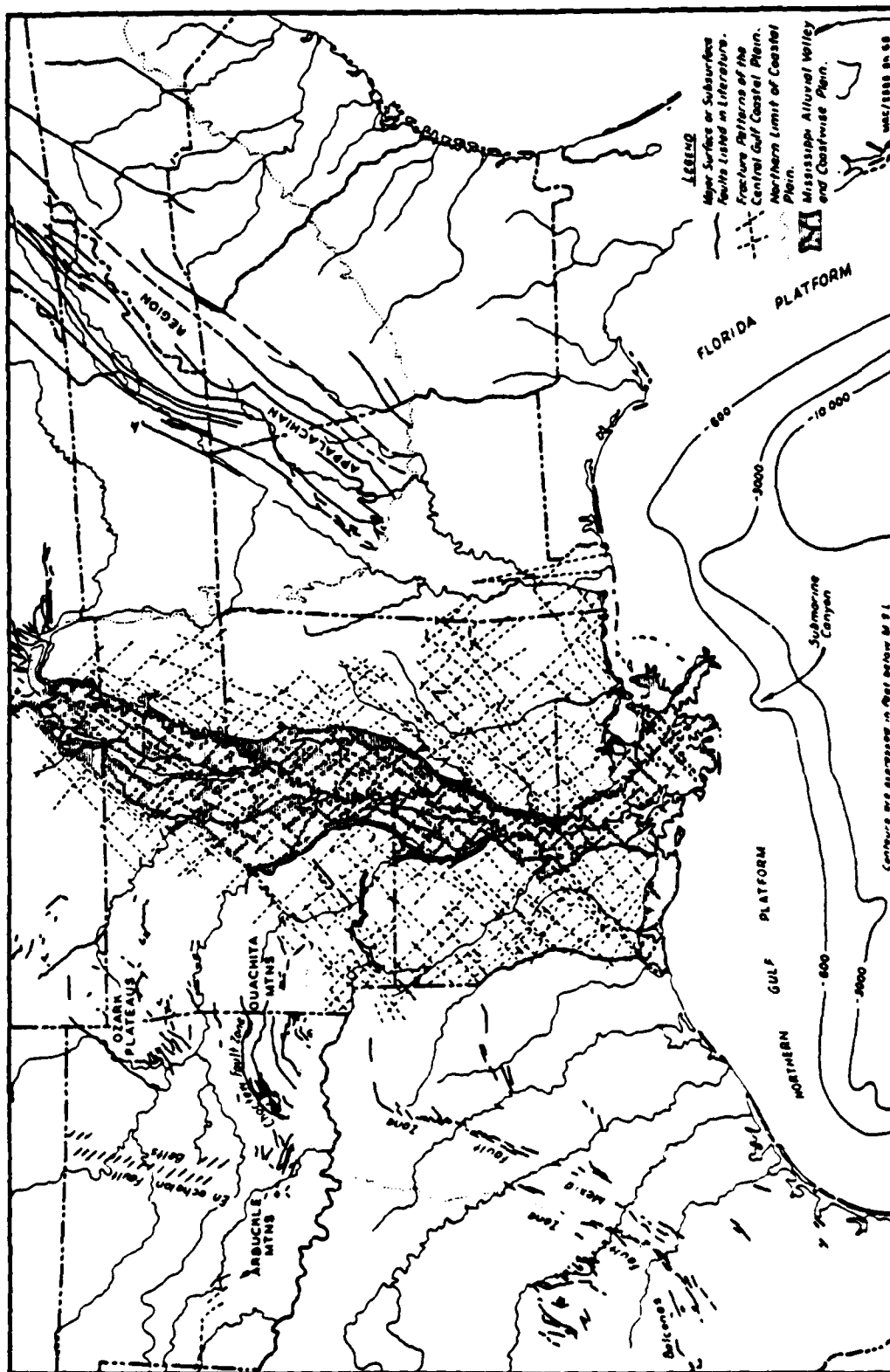


Figure 4.4 Regional fracture pattern of southern United States. (Fisk, 1944.)

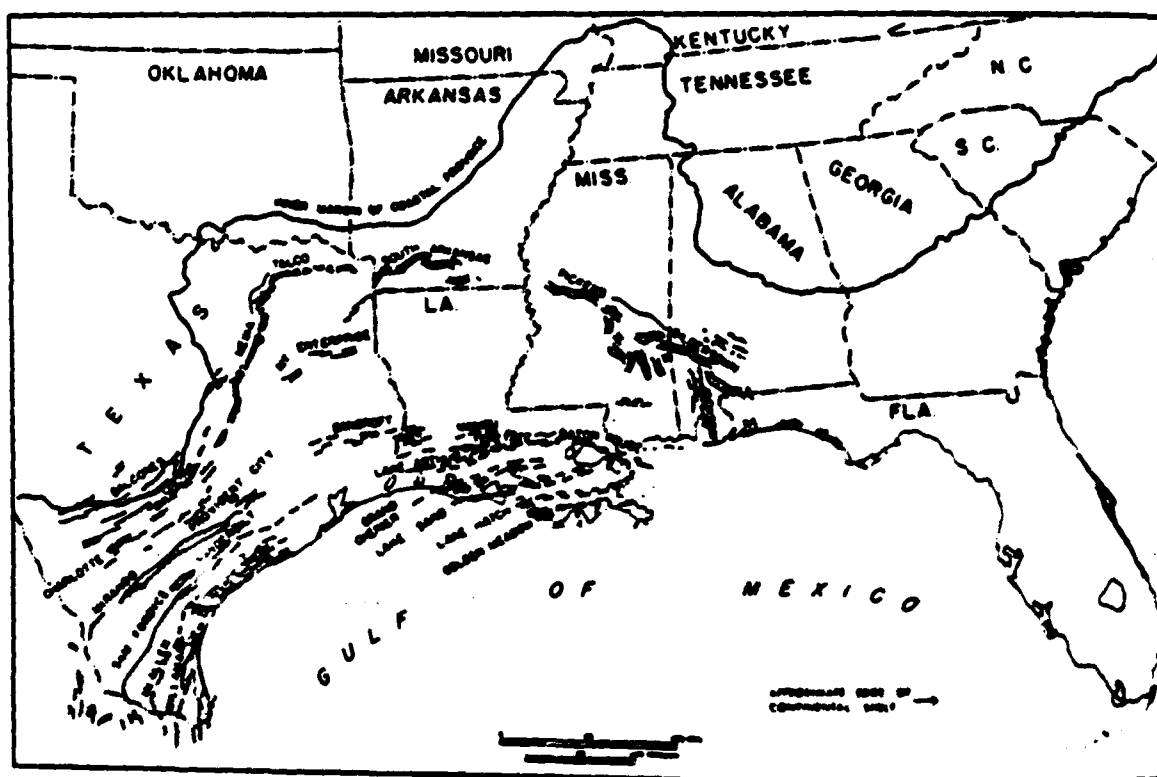


Figure 4.5 Diagrammatic representation of principal strike fault systems in northern Gulf coastal province.

subsidence, it was the locus of alkaline intrusions and is the site of high seismic activity (Hinze, et al., 1977). The embayment has been interpreted as a rift or failed arm about a triple junction near Jackson, Mississippi. Another possibility is that the embayment is located along an old zone of weakness that controlled the location of a major fracture zone farther to the south, and it was active in the opening of the Gulf of Mexico. A preexisting rift zone appears to have been reactivated to form the present embayment. It is difficult to choose between the above hypotheses; nevertheless, seismic activity near the head of the embayment clearly does not occur randomly, but it is somehow related to the development of the embayment and to the emplacement of alkaline igneous rocks.

With the exception of normal growth faulting on the Gulf coastal plain, the central eastern United States is now dominated by compressional tectonics (Zoback and Zoback, 1980). Unlike the Gulf Coast area, the central mid-continent region of the United States in the vicinity of the confluence of the Ohio and Mississippi Rivers has long been regarded as part of the stable craton (Hinze, et al., 1977). Surficial and geological evidence has led observers to the assumption that during the past several hundred million years this area has undergone only minor tectonism which took the form of broad slow events. However, recent geologic evidence indicates that the mid-continent area has been and is currently tectonically active. This conclusion results from studies of earthquake epicenters in the mid-continent and other geophysical studies. Certainly, continuing geological and geophysical investigations show evidence of a complex structural development of the area. A knowledge of these structures is limited because of their subtle

nature, their masking by younger sediments and sedimentary rocks, and because of the limited amount of deep drilling.

Contemporary geodynamics of the central mid-continent of the United States is poorly understood at present, and the folding and earthquake activity in the New Madrid area has been considered anomalous, as the area is part of a plate interior.

Earthquake activity provides an important source of information about recent tectonism in the central mid-continent; however, the historical record is less than 200 years long. Tectonic implications from seismicity are hampered due to the low level of seismicity relative to active earthquake zones such as the western United States. The most intense earthquake activity has been centered in the New Madrid seismic zone of southeastern Missouri and adjacent areas. The historical earthquake record in this area is dominated by the 1811-1812 earthquake sequence. The three major shocks of the 1811-1812 sequence had magnitudes of 7.2, 7.1 and 7.4. Recently detailed seismic information has become available through operation of a microearthquake array. The pattern of seismicity during twenty-one months of recording has revealed several linear trends in northeast-southwest and northwest-southeast directions (Fig. 4-6). These trends are interpreted as indications of the pattern and extent of present day active faults. An analysis of precise leveling data provides an important source of information of vertical movements of the earth's crust. A detailed analysis of vertical crustal movements in the eastern United States by Brown and Oliver (1976) shows that modern vertical movements appear to be related to earlier Phanerozoic trends. However, the rates of modern movements are much larger than average rates from the last hundred and thirty million years.

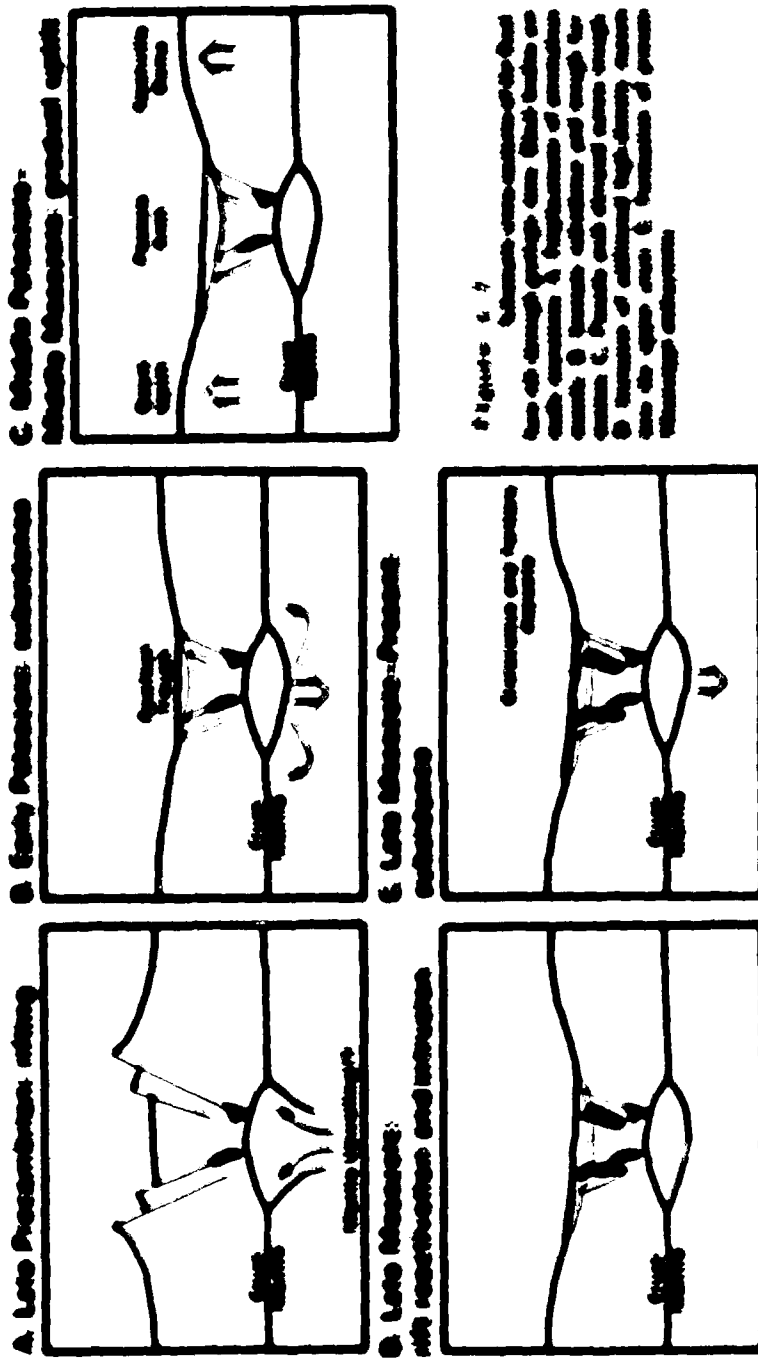


Figure 4.6 Map of the northern Wisconsin and southern Minnesota region showing earthquake epicenters (open circles with 1/2" diameter), plateaus (shaded areas), and all boundaries (heavy solid lines, 1/4" and 1/8" thick) and faults (thin solid lines, 1/8" thick).

Thus, the modern movements must be episodic or oscillatory.

The most significant observations on vertical movements in the mid-continent area are an indication of a general eastward tilt of the mid-continent and a southward tilt of the Gulf Coastal Plain including the Mississippi embayment (Brown and Oliver, 1976).

(Evitt and McGinnis (1976) conclude that the Mississippi embayment is the result of reactivation of the Reelfoot rift in Mesozoic time. Rifting has clearly played an important role in tectonic development of eastern North America. Renewed movement of old rift-valley fraction can be expected. The structural history of the Reelfoot rift is summarized by (Evitt and McGinnis (1976) is presented on Figure 8.7 and as follows. Geophysical and geological data indicate that the Mississippi embayment is a site of former continental rift, herein called the Reelfoot rift. The arching of the crust over a thick linear body of mantle as interpreted from geophysical data suggests that several kilometers of uplift and erosion occurred (Fig. 8.7A). This rifting must have occurred during Precambrian time. Following failure of rift development isostatic forces caused subsidence of the rift that formed the Reelfoot basin (Fig. 8.7B), which was several kilometers deep. From middle Paleozoic time through middle Mesozoic time the rift area was subjected to only mild tectonism (Fig. 8.7C). By late Mesozoic time there was renewal of activity along the rift with emplacement of large plutons derived from partial melting of mantle (Fig. 8.7D). Isostatic adjustment due to the intrusion of these high density rocks into the upper crust promptly renewed subsidence, which is occurring at present time (Fig. 8.7E). The modern embayment is thus the product of a long sequence of geologic events that began with a failed rift in late Precambrian time.



Uplift also took place in parts of the dome during the 1811-1812 New Madrid earthquakes. Russ (1961) concludes that the Fayetteville Dome is a product of at least three distinct seismically generated pulses of uplift as well as an unknown component of eustatic deformation. The timing of deformation on Bigley Stage is less clear than that of the Fayetteville Dome. Russ (1961) suggests, however, that the uplift is out of phase with and occurred earlier than the uplift in the dome.

Recent information of the Lake County uplift near Fortageville and Linden, Missouri and of part of Fayetteville Dome is supported by several lines of evidence. For example, landlocks used to regularly make passage from the Mississippi River through Portage Bayou past the present town of Fortageville and on to St. Louis River following the 1811-1812 earthquakes, such passage was no longer possible because the dome apparently had been uplifted. The upwarping probably occurred in an area southeast of Portage Bayou, Missouri along the northwestern edge of the Fayetteville Dome, and it may have formed a waterfall that reportedly formed across the Mississippi River during the earthquake of February 7, 1812.

CHAPTER 5

DEFORMATION OF THE ALLUVIAL VALLEY

This chapter provides analyses of three major tectonically active features in the Mississippi alluvial valley: the Late County Uplift in the upper valley, the Monroe Uplift in the middle valley, the Higgins Uplift in the lower valley. These three major features, or groups of features as in the case of the Higgins Uplift, were selected to provide evidence of the effects of neotectonic activity throughout the geographical limits of the valley, not just at one latitude.

Upper Valley: Late County Uplift

The great 1811-1812 earthquakes near New Madrid, Missouri have created considerable concern about the possibility of a repetition. Therefore, extensive studies have been carried out in this area by agencies of the federal government in cooperation with scientists at universities. Literature relating to the geophysics and geology of the Mississippi embayment between Memphis and Cairo is abundant. There are two papers that summarize this work, one by Stearns (1979) and the other by Ross (1981). They draw upon the work of others and their own research to conclude that there is active tectonism in the upper valley, and the following is a condensation of their work, as it is related to the objectives of this project.

An area of deformation has been recognized near New Madrid, which is referred to as the Late County Uplift (Fig. 5-1). The surface of the uplift is as much as ten meters above the general level of the Mississippi River valley. The deformed area has a maximum length of about 50 km and the maximum width of about 23 km. Its relief is uneven

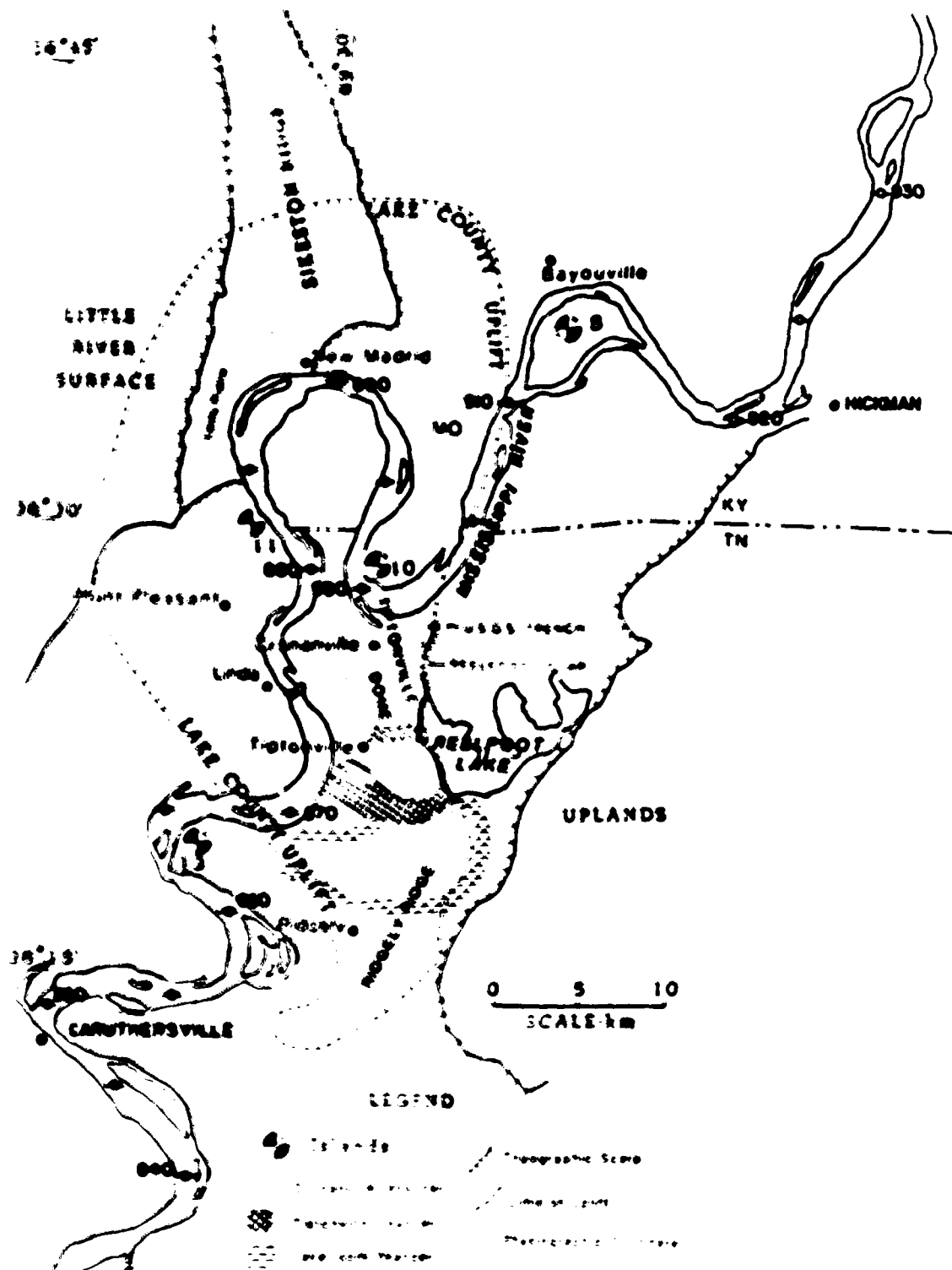


Figure 91 Location map of Lake County Uplift
(From Ross, 1982)

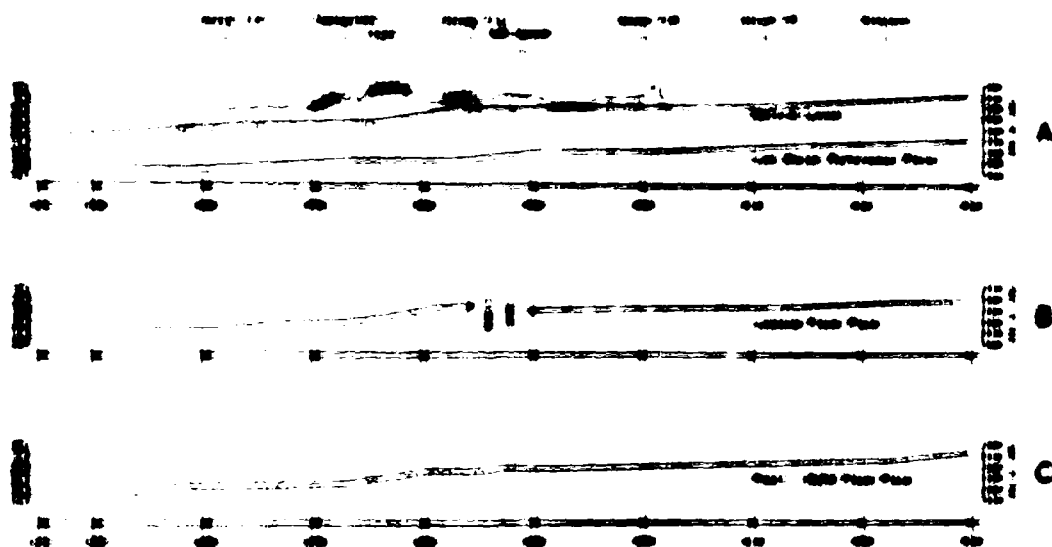


Figure 1. The effect of the concentration of the solution on the rate of the reaction. The rate of the reaction was measured by the change in the optical density of the solution at 440 mμ. The concentration of the solution was varied from 0.01 to 0.1 M. The rate of the reaction increased with increasing concentration of the solution. The rate of the reaction was also affected by the temperature of the solution. The rate of the reaction increased with increasing temperature of the solution. The rate of the reaction was also affected by the pH of the solution. The rate of the reaction increased with increasing pH of the solution.

of the Lake County uplift reveal that the structure is significantly higher than the natural occurring landforms of the modern meander belt (Fig. 5-2A) and the modern flood plain is also warped (Fig. 5-2b, c). 2) The longitudinal profiles of old abandoned river channels and natural levees that are related to an older meander belt show conclusively that the original channel and levees have been significantly warped, some to the extent that the original river flow direction has been reversed (Fig. 5-2a). 3) Reelfoot scarp vertically offsets abandoned Mississippi River channels which once flowed across the area. 4) The structures identified in an exploratory trench that was excavated across the Reelfoot scarp show that the scarp is a monoclinal structure formed as the adjacent Tiptonville Dome was uplifted. 5) Late Cretaceous, Paleozoic and Tertiary rocks are arched upwards over an area generally coincident with the greatest uplift on the Tiptonville Dome and Ridgely ridge. Beneath the uplift the post-Paleozoic rocks are generally characterized by intense deformation, that includes local doming, fracturing and faulting; apparent igneous intrusive masses have also been identified.

Russ states that although the earliest workers believed that the Lake County Uplift was deformed during the New Madrid earthquake, a significant amount of evidence suggests that the deformation has occurred in spacially irregular pulses for a period of at least several thousand years. The oldest landform to be affected by the uplift is Sikeston ridge, an early Wisconsin braided stream surface, that probably formed more than thirty thousand years ago (Fig. 5-1).

Radiocarbon dating shows that the deformation of the Tiptonville Dome has occurred within the last two thousand years, and structural

evidence observed in the trench indicates that most of the uplift that produced the Reelfoot scarp and the adjacent Tiptonville Dome occurred during two pre-1800 high intensity earthquakes.

An examination of Figure 5-2 reveals that all of the profiles have a similar shape suggesting that they may be the result of the same events. The profiles are convex upward, a configuration that is commonly associated with uplift. As delineated on Figure 5-2A, the natural levee gradient north of Island Number 8 (Fig. 5-1) is 0.33m/km, an average value for most of the natural levees in the upper Mississippi embayment. South of Island Number 8 the natural levee gradient decreases to 0.19m/km. The flattening is greater than that which normally occurs in a high sinuosity reach of the Mississippi River, and it is probably the result of upwarping along the northern edge of Tiptonville Dome. In the vicinity of Island Number 11 the natural levee gradient steepens sharply for a distance of 11km. This reach is in the approximate location of one of the two waterfalls that were reported to have formed on February 7, 1812. It is possible that the faulting that produced the waterfall also caused uplift and tilting of the adjacent floodplain. Between Linda and Island Number 13 the natural levee gradient flattens again. Here the flattening occurs in a reach where the river flows directly down the regional slope. Such a gradient would not normally be expected to occur as a result of fluvial process, and it is possible that minor uplifts deform this section of the river also. Near Island Number 13 the river leaves the vicinity of the Lake County uplift, and the natural levee has a slope typical of most of the meander belt in the upper embayment.

Russ points out that the Mississippi floodplain, (Fig. 5-2c) which has formed since 1820, has a profile that closely matches those of the

lowland floodplain and natural levee. He states that this may be because recent and even current deformation is warping the modern floodplain surface. Supporting evidence for the latter interpretation is found on the low water reference plain (Fig. 5-2a) which has a subtle, but noticeable convex upward profile. This is good evidence that significant modification of floodplain is occurring. Russ also states that several aspects of the meander pattern of the Mississippi River suggest control by tectonic processes. Between Cairo, Illinois and Hickman, Kentucky, the river is presently relatively straight. From Hickman south to Arkansas, however, it is highly sinuous (Fig. 5-1, 5-3). It is possible that the river straightened its course in order to increase the slope in an area where tilting is reducing it. Since the uplift of the Tiptonville Dome, the migration of the large New Madrid meander has undoubtedly been inhibited. The river is wedged between the topographically high Sikeston Ridge to the north and the Dome to the south.

The position of the Mississippi River course within its meander belt also suggests the possibility of tectonic influence. Between Cairo, Illinois and Hickman, Kentucky and in general between Blytheville, Arkansas and Memphis, Tennessee the river flows along the eastern edge of its meander belt (Fig. 5-1, 5-3). However, between Hickman and Blytheville the river shifts out into the western half of its meander belt. It is conceivable that the river has been deflected to the west as a result of the uplift. Old maps indicate that the position of the river in 1765 is similar to the river's modern location. Thus, any significant tectonic deflection must have occurred before 1765.

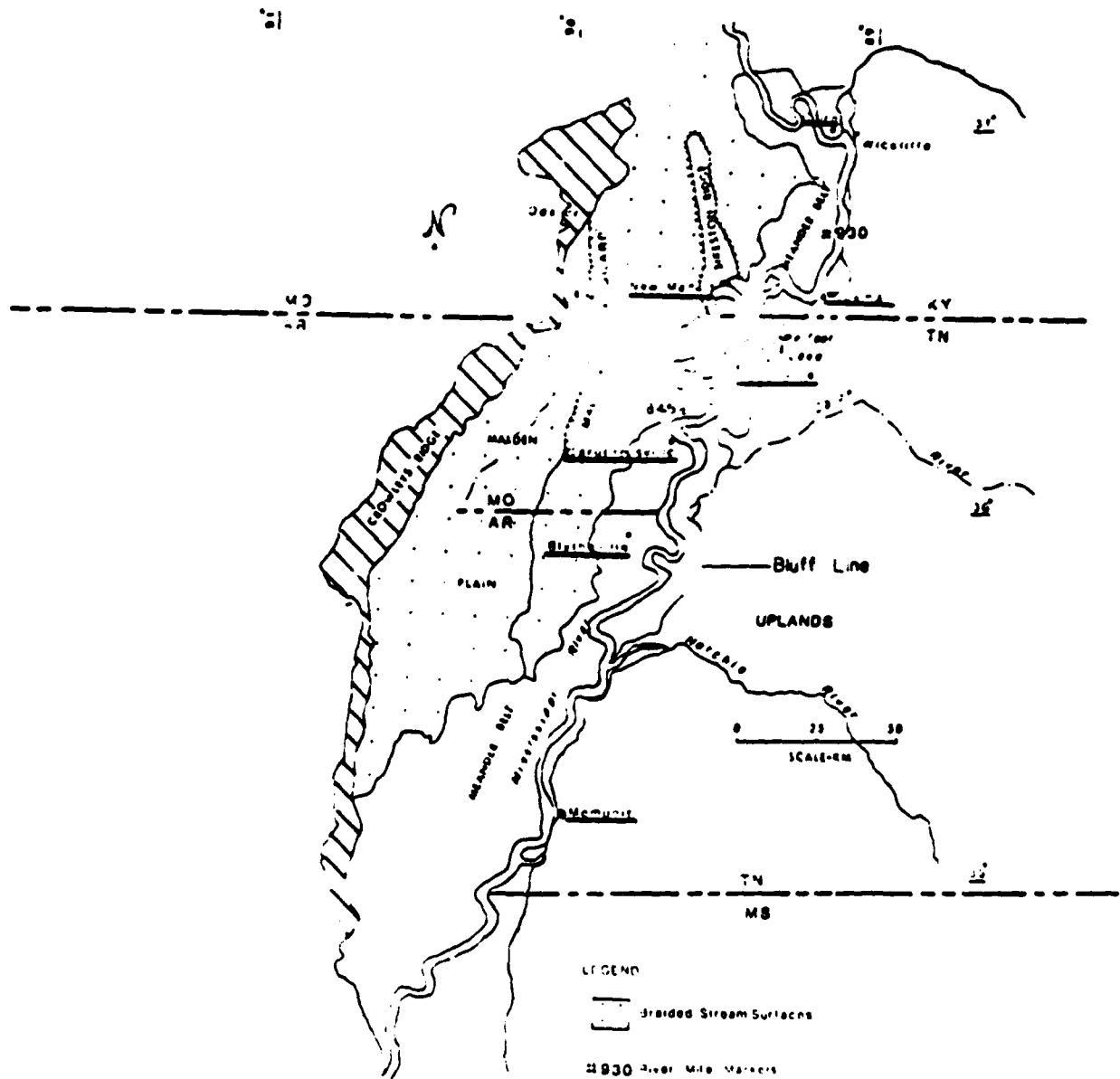


Figure 5.3 Mississippi Valley above Memphis.

The evidence indicates that the surface geologic structures in the New Madrid region are complex composite features that formed during a series of tectonic events. Most of the deformation was associated with large magnitude earthquakes. The geometry and location of the surficial structures are to a larger measure controlled by underlying ancient and deep seated geologic structures. Most of these structures lie along the axial zone of an inferred Precambrian crustal rift. The complexity of the Lake County uplift is revealed by the presence of several distinct topographic bulges within the zone of upwarping. The bulges exhibit differences in relief, geometry, orientation, time of formation in relation to subsurface structures, gravity and location of modern seismicity. The seismic studies by St. Louis University reveal that this region and an adjacent part of the Little River surface are sites of numerous recent micro earthquakes (Fig. 5-4). It is probable that stresses that produce the uplift are still active today.

Walters (1975) tried to evaluate the effect of the New Madrid earthquake on the river downstream of the uplift. He reviewed the descriptions of the river following the earthquake, and he concluded that the segment of river beginning some distance upstream of Island Number 10 down to about Island Number 32 suffered the most severe damage, especially by extensive bank failure. There is evidence that the width increased following the earthquake in a reach from Cairo to Island Number 32 near Osceola, Arkansas. He found that the mean bankfull width in 1821 was 3100 feet. Most of the width measurements from Cairo to Island Number 32 plot above the mean line, but downstream of Island Number 32 the width measurements begin to plot below the mean line except at large islands.

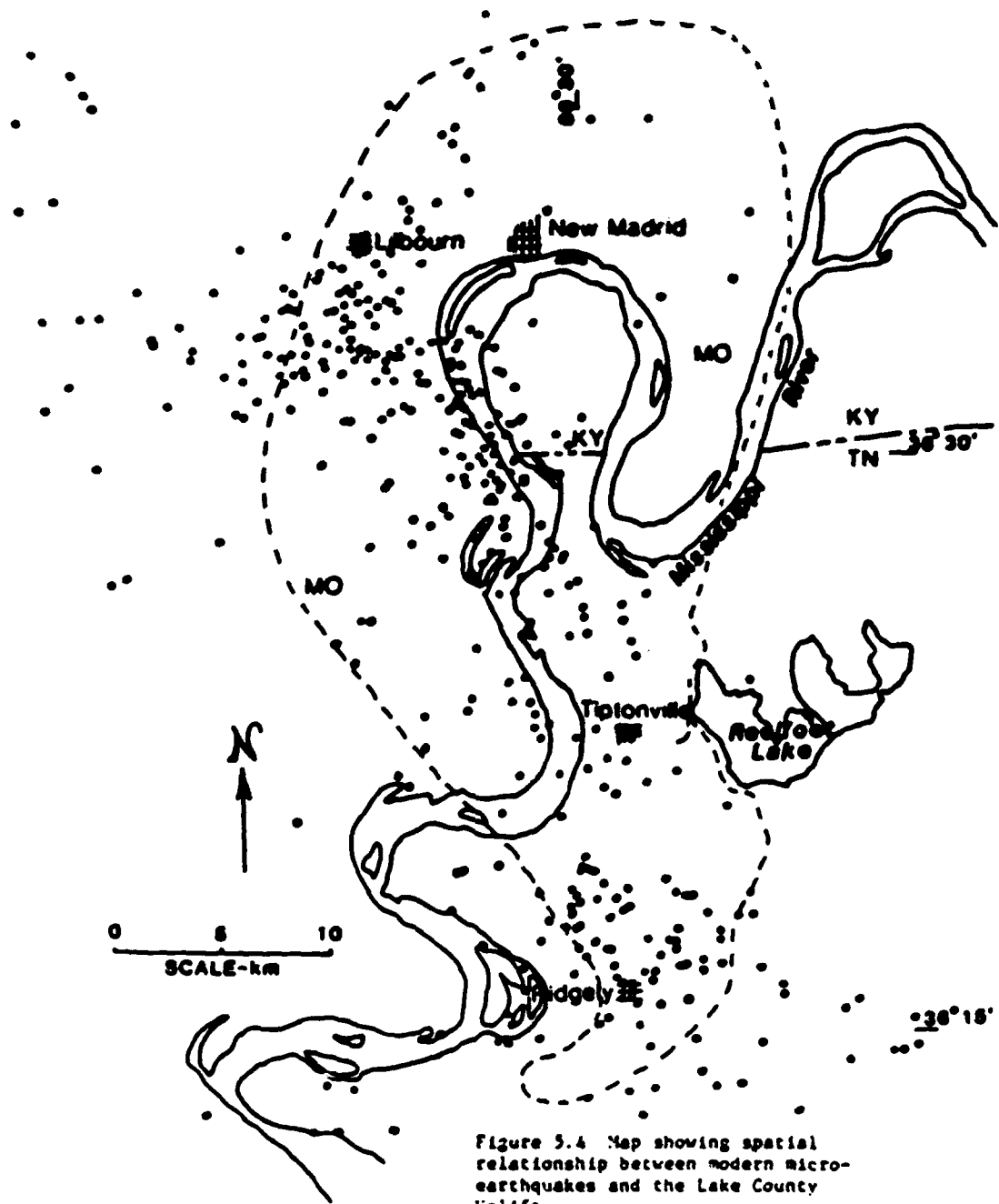


Figure 5.4 Map showing spatial relationship between modern micro-earthquakes and the Lake County Uplift.

From examination of gaging station records and especially specific-gage relations, Walters concludes that the lower Mississippi River channel from above New Madrid, Mo. to Red River Landing, La. was aggrading after about 1880 perhaps as a result of the sediment introduced by the New Madrid earthquakes. Walters considered that the influx of sediment into the channel caused a reduction in meandering by an increase in cutoffs. Figure 5-5 indicates the direction of change of river characteristics. Figure 5-6 shows that the frequency of cutoffs increased sometime after the earthquake of 1812. There was a peak period of neck cutoffs between 1818 and 1874. From 1875 to 1932 the number of neck cutoffs decreased considerably.

Walters summarizes as follows: From 1765 to the winter of 1811-1812 the lower Mississippi was a graded river. Cutoffs were not occurring at an unusual rate, and overbank flows were being temporarily stored and released by the St. Francis and the Yazoo basins. Beginning on December 16, 1811 and continuing intermittently through February, 1812 the New Madrid earthquake shocks caused bank caving which introduced tremendous quantities of sediment into the channel. The most severe caving occurred in the reach from the confluence of the Ohio and Mississippi Rivers to a few miles below what is now Osceola, Arkansas. The sediment increase caused excessive shoaling and an enlargement of the channel islands. At some locations new islands and point bars were formed. During the years following the earthquake, the excess sediment began to move gradually downstream to the lower reaches below Osceola. This increase in sediment load caused an increase in the meander rate, which is indicated by the fact that the vast majority of cutoffs occurred below the Cairo to Osceola reach. Above Osceola the

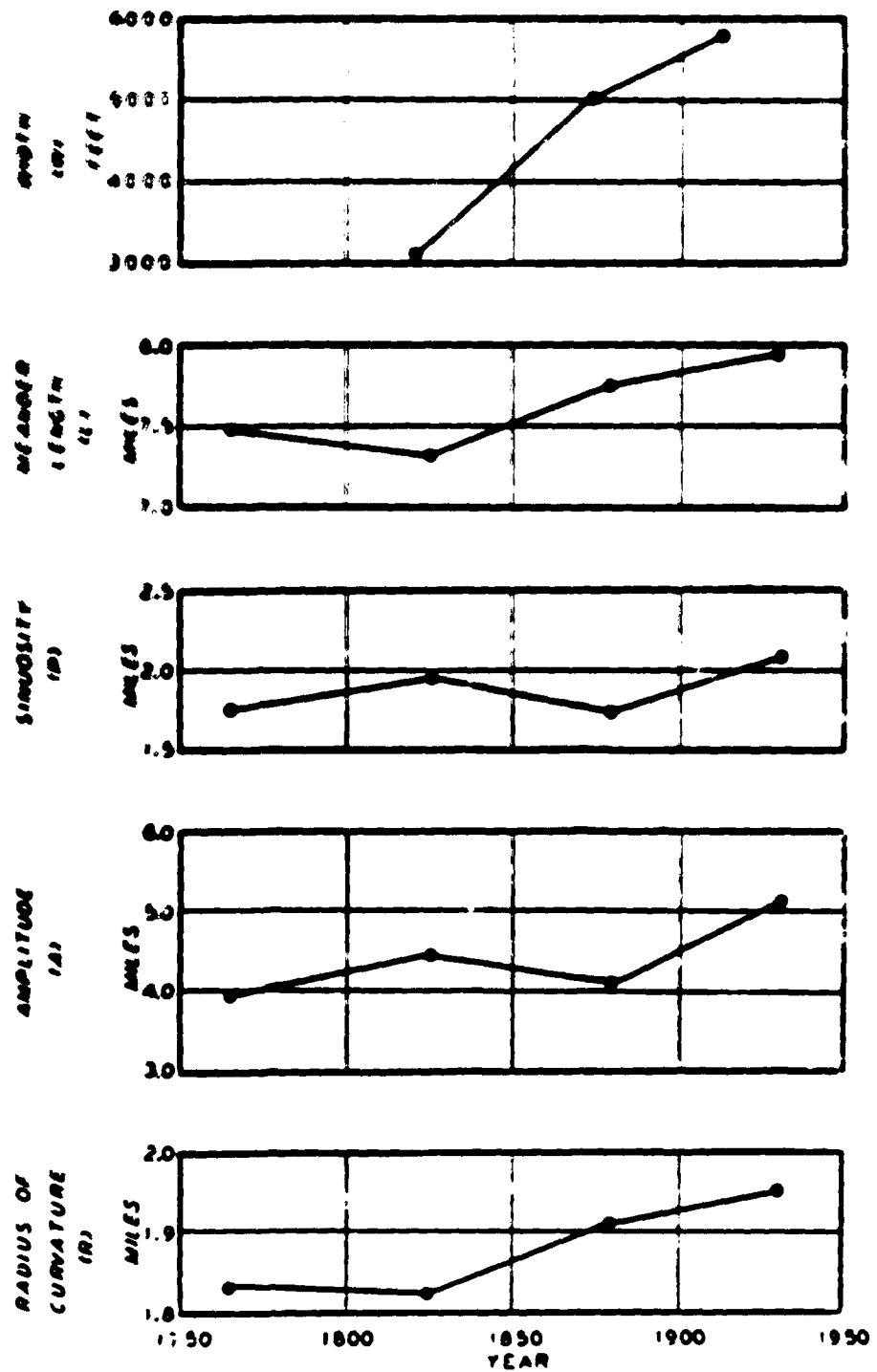


Figure 5.5 Direction of Change of Geometric Variables. (Walters, 1975)

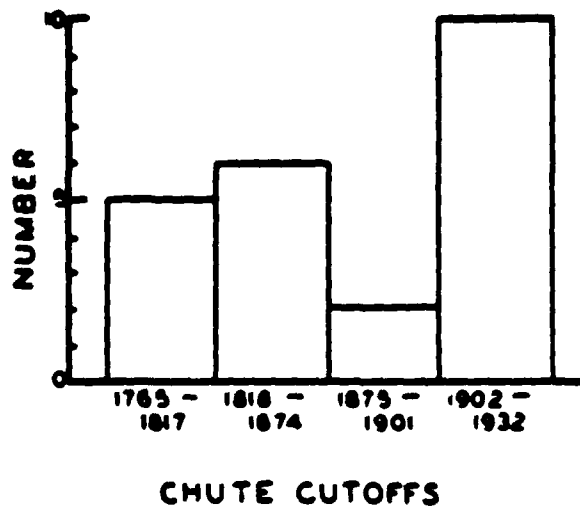
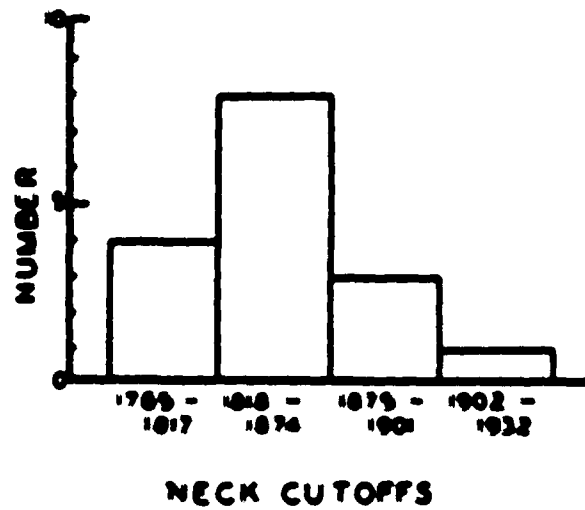


Figure 5.6 Occurrence of Cutoffs on the Lower Mississippi River from Cairo, Ill., to Red River Landing, La. (Walters, 1975)

introduction of sediment was almost instantaneous and the response was the formation of a wider aggrading channel. Below Osceola the response was a formation of cutoffs.

The increase in sediment supply could be due to deforestation and land use change in the Mississippi River Basin. The extent of increased sediment supply due to land use change or to the New Madrid earthquake series is difficult to access.

A major problem in deciphering the effects of earth movements on streams is that man's activities have caused such major changes that disguise neotectonic effects. Recent channel changes in the Mississippi valley were caused, primarily, by man as a result of channel shortening, dredging, and levee building (Winkley, 1977).

Middle Valley - Monroe Uplift

THE MONROE UPLIFT

The Monroe Uplift, the extent of which is defined by deformed Cretaceous and Tertiary strata, is a dome approximately 75 miles in diameter. It is situated mostly in northeastern Louisiana, (Fig. 5-7, 5-8), but it extends into southeastern Arkansas and west central Mississippi (Wang, 1952). Its western boundary is defined by a southeast trending fault zone which runs just west of the city of Monroe, Louisiana (Easton, 1974). The southern edge of the uplift lies just north of Winnsboro. Its eastern-most extension includes the Mississippi River between Greenville and Vicksburg.

The uplift consists of the Boeuf and Tensas basins as well as Macon Ridge and the Mississippi River from mile 450 to 530. The basins are composed of large backswamp areas crossed by several old Arkansas River channels and well developed natural levees. The northwestern portion of the uplift includes part of the Western Highlands, and the southwestern

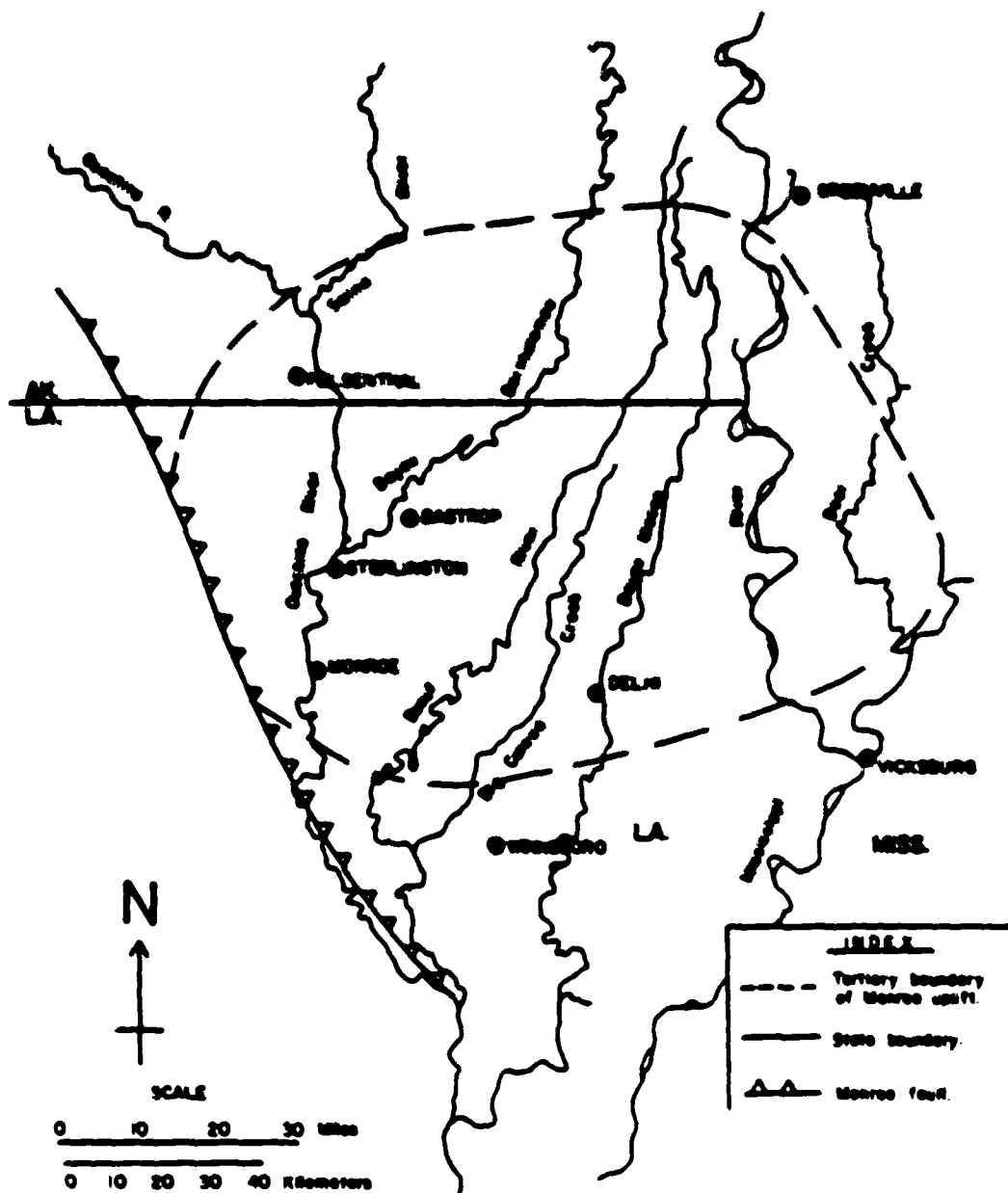
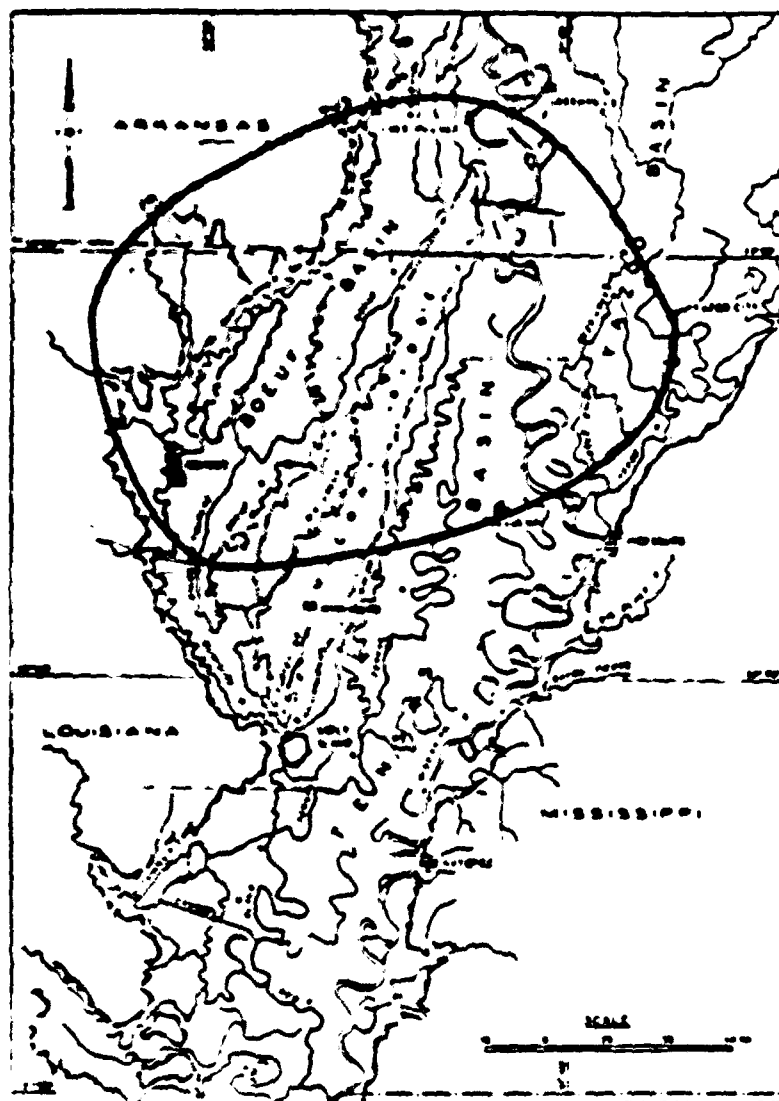


Figure 37



Geographic setting, Macon Ridge and vicinity.

Figure 5.3 Index map of Monroe Uplift area.
(from Saucier, 1968)

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INVESTIGATION OF NEOTECTONIC ACTIVITY WITHIN THE LOWER
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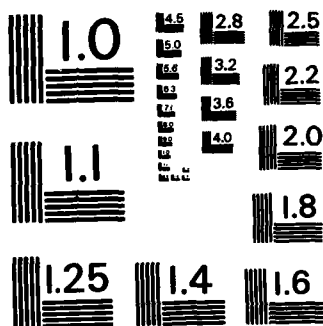
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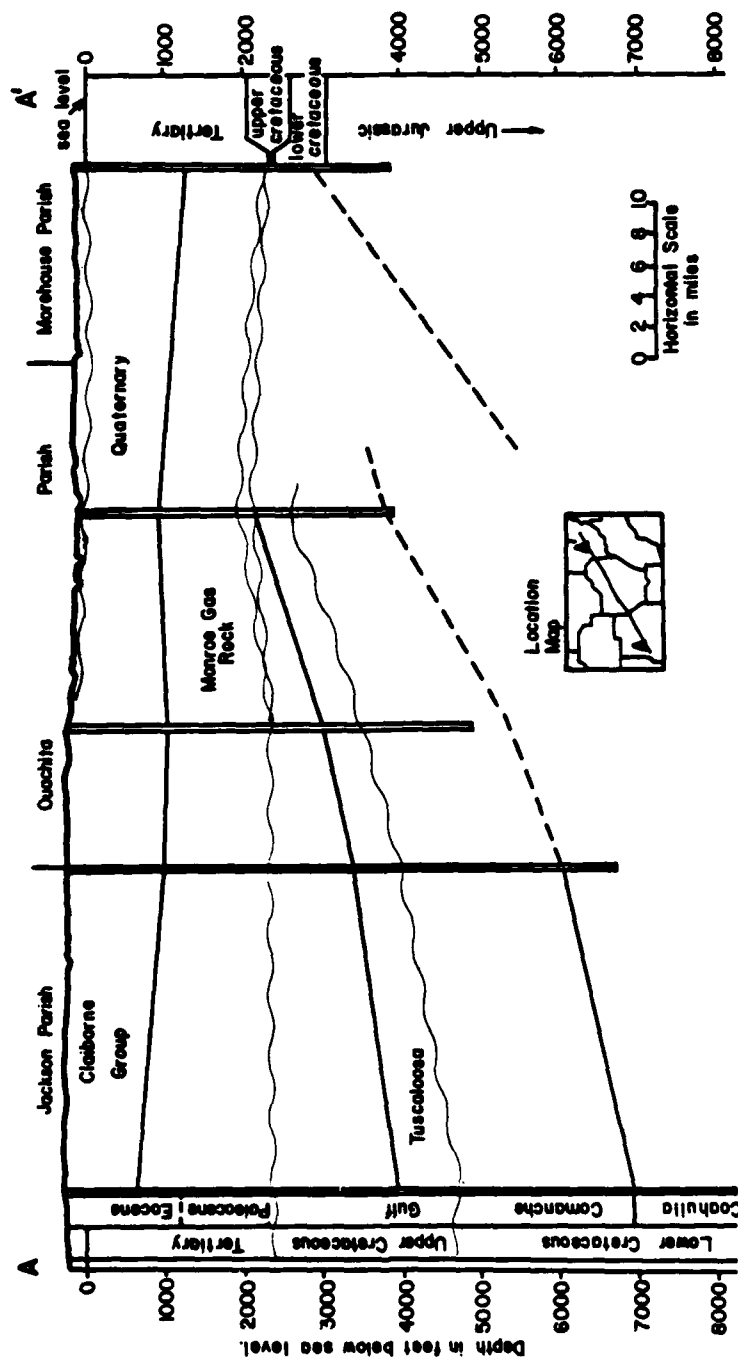
edge is defined by the prominent Highlands scarp. The Macon Ridge in the center of the Uplift is composed of five mid-Wisconsin glacial outwash terraces (Saucier and Fleetwood, 1970), which form a narrow elongated area of high ground which trends north-south.

A means of differentiating man's effects from neotectonic effects is to examine streams in a structurally active area that has not been appreciably affected by man. Therefore, several streams, which cross the Monroe Uplift in northeastern Louisiana and which generally parallel the Mississippi River were selected for study. The relatively small size and low energy of these streams, as compared to the Mississippi, provide an additional advantage because they should show the clearest effects of neotectonics due to their slower rate of adjustment to change.

The present streams flow generally to the southwest across the uplift and they locally occupy old abandoned courses of the Arkansas River. The streams which were examined, are (from west to east) the Ouachita River, Bayou Bartholomew, Boeuf River, Big Colewa Creek, Bayou Macon, and Deer Creek (Fig. 5-8).

The underlying stratigraphy of the area consists of over 10,000 feet of Mesozoic sand and clays overlying the pre-Cambrian basement rocks (Fig. 5-9). An unconformity separates the overlying Tertiary deposits from those of the Cretaceous age. Another major erosional unconformity forms the upper Tertiary boundary, and is overlain by a sequence of upward fining Pleistocene sands and gravels, which have filled deeply entrenched early Wisconsin age valleys.

The uppermost deposits are mid-Wisconsin braided-stream deposits that form Macon Ridge and 75 feet thick deposits of modern backswamp,



Stratigraphic section through Jackson, Ouachita, and Morehouse Parishes, Louisiana (from Wang 1952)
Figure 5.9

point bar and levee deposits in the basins. These modern stream deposits abruptly truncate the Pleistocene sand and gravels as the Macon Ridge deposits (Saucier, 1968).

UNDERLYING STRUCTURE

The structure of the Monroe Uplift is described as a complexly truncated dome, the oldest truncated rocks are Jurassic, the youngest, on the flanks, are Upper Cretaceous, and the truncating beds are Paleocene in age (Wang 1952). Deformation began as early as Jurassic time but the youngest rocks showing obvious tilting are those of upper Tertiary age.

Two periods of major uplift activity have been identified (Thomas 1950; p. 1504). The first occurred during upper Cretaceous time, and it is associated with intrusive and extrusive igneous activity. Then a period of rising sea level at the end of the Cretaceous caused deposition of sands and silts forming the Monroe gas rock. Subsequent uplift began in early Tertiary time.

Fisk (1939) states that the Jackson formation was tilted to the southeast by late Monroe uplift activity. Wang (1952; p. 76) suggests that the upwarping of the Monroe uplift continued through early Eocene time and that its axis gradually shifted south and southeastward. Thomas (1950; p. 1504) states that activity of the Monroe Uplift during Upper Cretaceous to Tertiary times has offset the synclinal axis of the Mississippi Embayment toward the west, in the area north of Humphreys County, Mississippi. He also suggests that uplift may have continued to the present.

Geologic evidence that the Monroe Uplift was active since the Tertiary has been presented by several authors. Veatch (1906) discussed the existence of two active linear structures, which pass through the

Monroe Uplift. These are the Angelina-Caldwell monoclinal flexure and Red River-Alabama Landing Fault (Fig. 5-10). According to Veatch, the Angelina-Caldwell flexure extends from Angelina County, Texas, through northern Louisiana, to the Mississippi River north of Vicksburg. Veatch claimed that there has been active upward tilting to the south along this linear zone since late Tertiary time. He further claimed that recent movement along the west end of the flexure has resulted in the formation of a series of shoals on the Sabine and Angelina Rivers and swamping of an area in the Angelina River valley in eastern Texas (Veatch, 1906).

Winkley (1980) discussed the possible effects of the Monroe Uplift on the past and present morphology of the Mississippi River in the vicinity of Greenville Bridge (mile 531.3). According to Winkley, the Monroe Uplift has upwarped the Tertiary formations below Greenville, Mississippi causing a pinching out the very resistant Yazoo Clay and exposing the easily erodable underlying Cockfield sand just downstream of Greenville Bridge. Figure 5-11 shows a NW-SE line which, according to Winkley, was the northern limit of the Monroe uplift.

Winkley states that the thalweg profile and the plan forms of the river, for both the current and historical channels, "change shape characteristics at the line of the Monroe Uplift" (pg. 36).

Fisk's (1944) maps of the old Mississippi River meander courses show that the Mississippi River has maintained very high sinuosity values above this line during the last 2000-6000 years. Meander loops have grown and cutoff in the same location several times but they have never migrated past the proposed line. In this sinuous zone, according to Winkley, the Mississippi River has adjusted laterally across the Yazoo clay, being unable to downcut through it. Below the line, Winkley (1980)

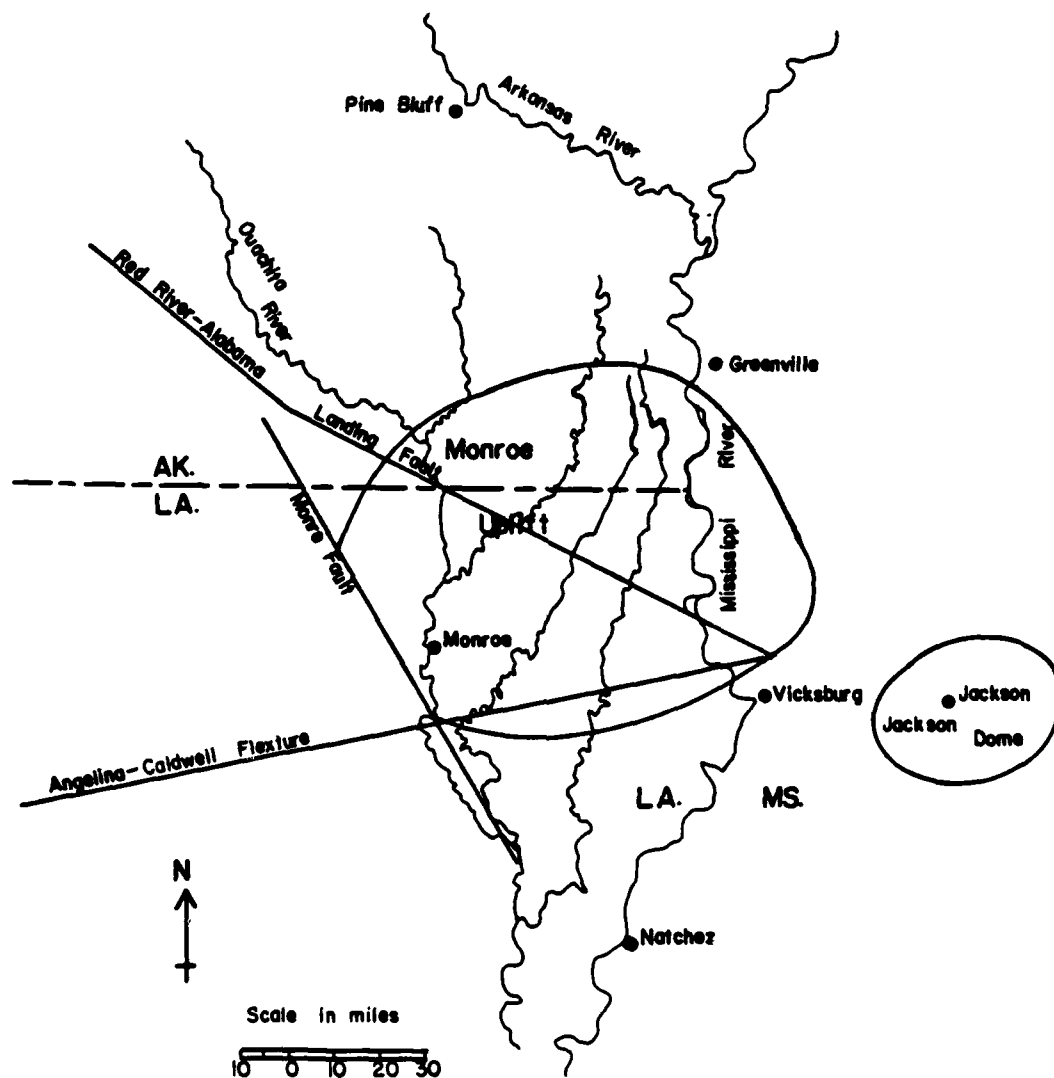


Figure 5.10 Geologic structures in the vicinity of the Monroe Uplift.

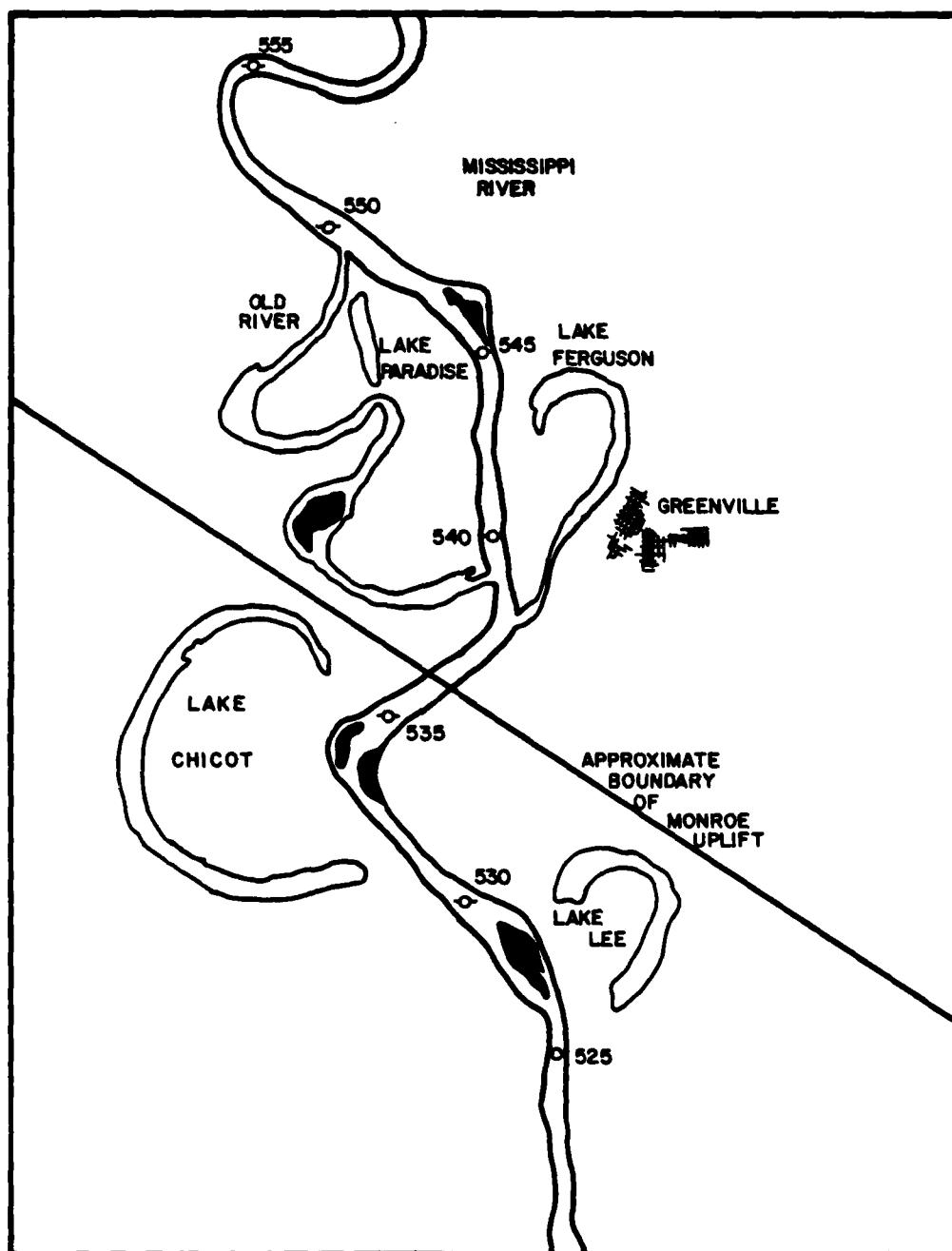


Figure 5J) Approximate boundary of Monroe Uplift near Greenville Bridge. (from Winkley, 1980)

states that the Mississippi River encountered the Cockfield formation, eroded a narrow thalweg and a deep plunge pool, and has persisted through time locking the river in place. Thus, according to Winkley, the post-Tertiary uplift caused entrenchment of the Mississippi River into the Cockfield sand through the uplifted area and consequent confinement of the river over the uplift, and it caused a highly sinuous reach of river above the uplifted zone in which the meander loops have actively migrated laterally across the resistant Yazoo clay but have never migrated downstream.

PLEISTOCENE AND HOLOCENE HISTORY

The Quaternary alluvial deposits in the Monroe Uplift area form a series of terraces and the present valley floor within the Mississippi embayment. These deposits are confined to the east and west by highlands of Eocene age, which rise abruptly 100 to 150 feet above the floodplain. The Eocene formations strike generally northeast and they dip gently to the southeast, due to regional tilting toward the Mississippi delta (Fisk 1944).

The Pleistocene and Holocene deposits in the area reflect three interglacial periods and two periods of Wisconsin glacial advance. (Saucier, 1968). The Sangamon Interglacial stage (around 60,000 BP) was characterized by a high stable sea level and deposition of the Prairie terrace sediments by mature meandering stream systems (Saucier 1968; Saucier and Fleetwood, 1970). The major fall of sea level associated with Early Wisconsin glaciation (40-60,000 BP) caused major entrenchment in the Mississippi Valley. Most of the Prairie terrace formation was removed (Saucier 1970), and a series of alluvial channels was formed in the Tertiary deposits deep within the valley

(Saucier, 1968).

An interglacial period followed during the Farmdalian Substage (about 33,000 BP) that was marked by a high sea level and rapid aggradation of the Mississippi Valley. The braided Arkansas River at this time formed a long narrow alluvial cone, the remnants of which now form Macon Ridge. Five terrace levels were formed and preserved on Macon Ridge as the Arkansas River shifted westward and slowly degraded due to falling sea level (Saucier and Fleetwood, 1970).

During the late Wisconsin regression sea level fell again thereby causing a second period of valley entrenchment which lowered the valley floor about 75 feet (Saucier, 1968). Subsequently sea level rose and reached present sea level about 3000-5000 BP. The result was a rise of base level and vertical accretion of the entrenched valleys up to their present elevation (Saucier, 1968).

EVIDENCE FOR RECENT ACTIVITY

The review of geometric effects of uplift in Chapters 1 and 2 shows that rivers will respond to uplift. Therefore studies of river morphology and anomalies in the area will indicate possible deformations. In addition, survey data provide information on relative surface changes.

SURVEYS

The most recent surface movements can be obtained by precise resurveys. Precise releveled routes in the vicinity of the Monroe Uplift area were resurveyed during a number of years. These locations include Monroe, Delta Point, Alexandria, Delhi, Winnsboro, and Columbia, Louisiana. The survey data indicate that relative subsidence of up to 160mm between 1934 and 1966 has occurred in the vicinity of Monroe.

However, an anomalously large lowering of the potentiometric surface also occurred in the Monroe vicinity. In the last 80 years the water level of wells piercing the Sparta Sand aquifer has decreased from +100 feet to as low as -160 below sea level (Ryals, 1980). These water level changes seem to correlate closely with the observed recent relative subsidence at Monroe. Therefore, evidence of present uplift activity as detected by resurveying may be complicated by the water level drops.

Therefore earth movements recorded by resurveying do not necessarily reflect long-term rates but only the most recent and probably temporary, subsidence rate. The subsidence of 160mm at Monroe is probably minor when compared to the long-term movements due to activity of the Monroe Uplift. Therefore, other methods had to be used to determine if recent earth surface movement had occurred in the Monroe Uplift area.

TERRACE PROFILE DEFORMATION

In order to identify displacement of a fluvial terrace by earth movements, the original geometry of the terrace must be known (Machida, 1960). The original geometry of a fluvial terrace can be assumed to be smooth and to decrease in slope in a downstream direction. It should generally parallel the modern stream gradients.

In this study several terrace plain profiles from the Monroe Uplift area have been analyzed in order to determine if they are deformed.

PRAIRIE TERRACE

The Prairie terrace (Fisk, 1939) is an extensive, easily identified Quaternary terrace, which is preserved along the margins of the Ouachita River valley and Red River valley. Small remnants of this surface occur on the eastern margin of Macon Ridge (Saucier and Fleetwood, 1970;

Saucier, 1967). The Prairie terrace formed during the Sangamon Interglacial Stage (Saucier, 1968). The preserved surfaces have been accurately delineated on surficial geology maps of the area (Saucier, 1967; Fleetwood, 1969; Smith and Russ, 1974). Fig. 5-12 is a generalized reconstructed longitudinal profile of the Prairie terrace along the Ouachita River valley and lower Red River valley. Also shown are the elevations of those remnants of the terrace preserved on Macon Ridge. The surface on the southern end of Macon Ridge is about 15 feet higher than the general profile elevation at the similar longitudinal distance along the Mississippi River valley. Since this remnant surface is located within the Monroe uplift, whereas the other surfaces are located along the western margins of the uplift, Quaternary activity of the Monroe uplift could explain such a vertical offset.

MACON RIDGE

The Macon Ridge (Figure 5-8) runs generally NNE to SSW through the Monroe Uplift, and it is composed of braided river deposits of the ancestral Arkansas River which were deposited during the Farmdalian Substage and at the beginning of the late Wisconsin regression (Fig. 5-12). Five terrace levels have been identified (Saucier 1969), and their present reconstructed profiles are shown in Figure 5-13. According to Saucier and Fleetwood (1970), the highest terrace (Qtd_1) was formed during the period of maximum sea level in the Farmdalian Substage; and the lower terraces (Qtd_2 - Qtd_5) were formed during the initial stages of the Late Wisconsin glaciation.

Since the Arkansas River was braided during formation of the Macon Ridge terraces (Saucier and Fleetwood, 1970), its sinuosity would be near unity. The stream profile of the river during formation of each of the

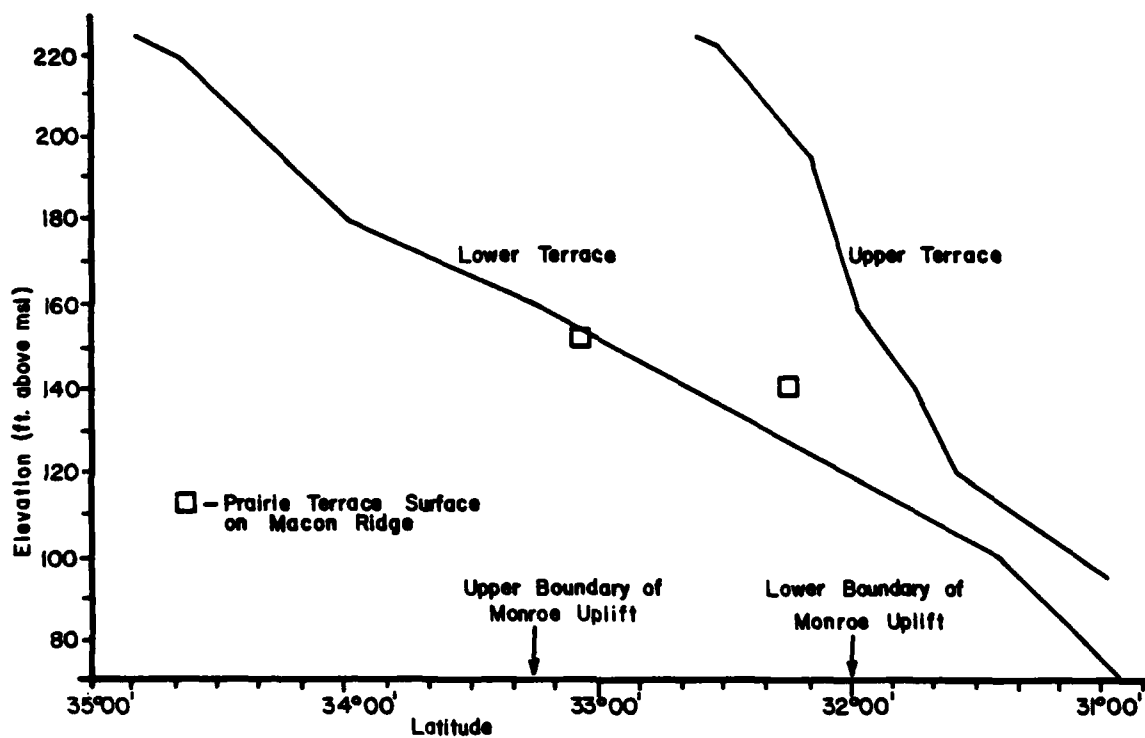


Figure 5.12

Longitudinal Profiles of Prairie Terraces along Mississippi River Valley.

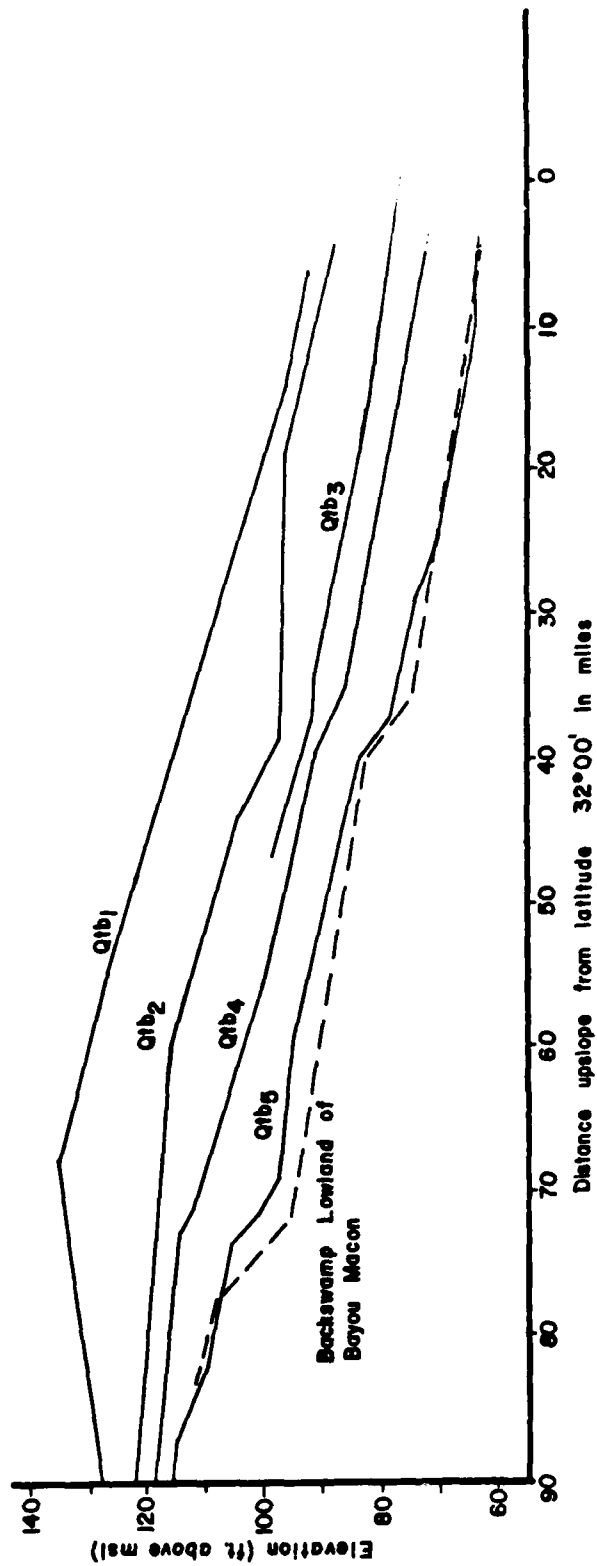


Figure 5.13 Surface Profiles of Macon Ridge Terraces and Backswamp Lowland of Bayou Macon.
Profiles taken parallel to section A-A' in fig. 5.9

terrace surfaces should be similar to the original profile of the terrace. Therefore, it is assumed that the original profiles of the lower Macon Ridge terraces were straight to convex-up in shape.

Analysis of the present terrace profiles of Macon Ridge (Figure 5-13) shows that there are several irregularities in the profiles that are not consistent with the original profiles of the terraces.

The most obvious irregularity is the abnormally large convexity displayed by Qtb_1 in which the terrace has a reversed profile slope north of mile 70. The Qtb_1 profile shows the largest convexity of all the terraces and it has clearly been deformed by uplift in the vicinity of mile 70 since its formation about 33,000 years ago. Downslope of mile 70 the terrace is steep, but upslope of mile 70 its gradient has become less through time.

The lower terraces do not show as much convexity but they do contain numerous anomalous breaks in their profiles. Between miles 32 and 46, short sections showing major increases in profile slope occur on the profiles of Qtb_2 , Qtb_3 , Qtb_4 , Qtb_5 as well as on the present floodplain and natural levee profile of Bayou Macon which parallels Macon Ridge (Fig. 5-13). Of particular note is the close parallelism between the profile of the lowest terrace, Qtb_5 and that of the present floodplain of Bayou Macon. These two surfaces are approximately 30 miles apart, in different basins, and they are separated by the higher terraces of Macon Ridge. However, both surfaces show breaks in their profiles at mile 76 and also at mile 40. Above the slope breaks, the slopes of both profiles are approximately 0.4 to 0.7 foot per mile. Then just downslope, at the breaks, there is a major increase of slope. For example, at mile 76 there is an increase in slope from about 0.6 ft/mi.

to 2.5 ft/mi. Downslope of the break, the slopes decrease to between 0.4 and 0.8 ft/mi. At the second break at mile 40 there is an increase in slope from 0.4 ft/mi. to about 1.2 ft/mi and then a decrease downslope to about 0.6 ft/mi.

From the distribution of the slope breaks it appears that between miles 40 and 76 the land surface has tilted to the north thereby reducing the gradients. The breaks themselves show between 5 and 10 feet of vertical displacement of the profile surfaces. Similar recent displacement may be occurring at depth along faults related to the Monroe uplift.

Because the terraces were formed at different levels at different times, rejuvenation of a river system alone cannot explain the parallel nature of the slope changes among the different terraces. There is no major change in lithology along any of the terraces, which would explain such breaks. Also the breaks in profile do not correspond to any major tributary confluences either in the past or present. Therefore, a reasonable explanation for these offsets is that they are surface expressions of vertical movement along underlying faults. The irregularities in the profiles of the Macon Ridge terraces and the Bayou Macon floodplain are all located within the boundaries of the Monroe Uplift as defined by Tertiary deformation. Therefore, if such faults do exist, then they most likely are related to recent movement of the Monroe Uplift.

Comparing the location of possible deformation of the oldest terrace with that of the more recent terraces (Fig. 5-13) it is possible that the type and location of surface deformation due to surface deformation due to deep-seated movement of the Monroe Uplift may be

variable through time.

During the formation of the oldest terrace (Qtb_1) uplift was concentrated at the northern end of the Monroe Uplift. The younger surface profile irregularities seem to indicate that the recent activity has shifted south.

DEWEYVILLE TERRACE

The Deweyville terrace is intermediate in elevation between the Prairie terrace and the Holocene floodplain, and it occurs as well preserved surfaces along the margins of the Ouachita River valley from central Arkansas to just south of Monroe, Louisiana (Saucier and Fleetwood, 1970). The terrace has been dated as mid- to late-Wisconsin (30,000 to 13,000 BP), and three distinct terrace levels have been identified (Saucier and Fleetwood, 1976). Detailed maps showing the specific localities and elevations of the three levels have been made by Fleetwood (1969) and Saucier (1967), and Saucier and Fleetwood (1970) produced and analyzed downvalley terrace profiles of the three levels.

The Deweyville surfaces were reexamined and longitudinal profiles were plotted using the surficial geology maps of Fleetwood (1969) and Saucier (1967) (Figs. 5-14, 5-15). An obvious convexity occurs on the highest terrace surface (Qtd_1) between 30 and 100 valley miles below Camden, Arkansas. Likewise, the intermediate terrace (Qtd_2) profile shows a possible convexity between valley miles 40 and 70. The location of the upper level profile convexity correlates very closely with the location and extent of the Monroe uplift.

BEACH RIDGE

A series of low ridges occurs on the Deweyville terrace at its contact with the Qtd_2 terrace. They have been delineated by Saucier and

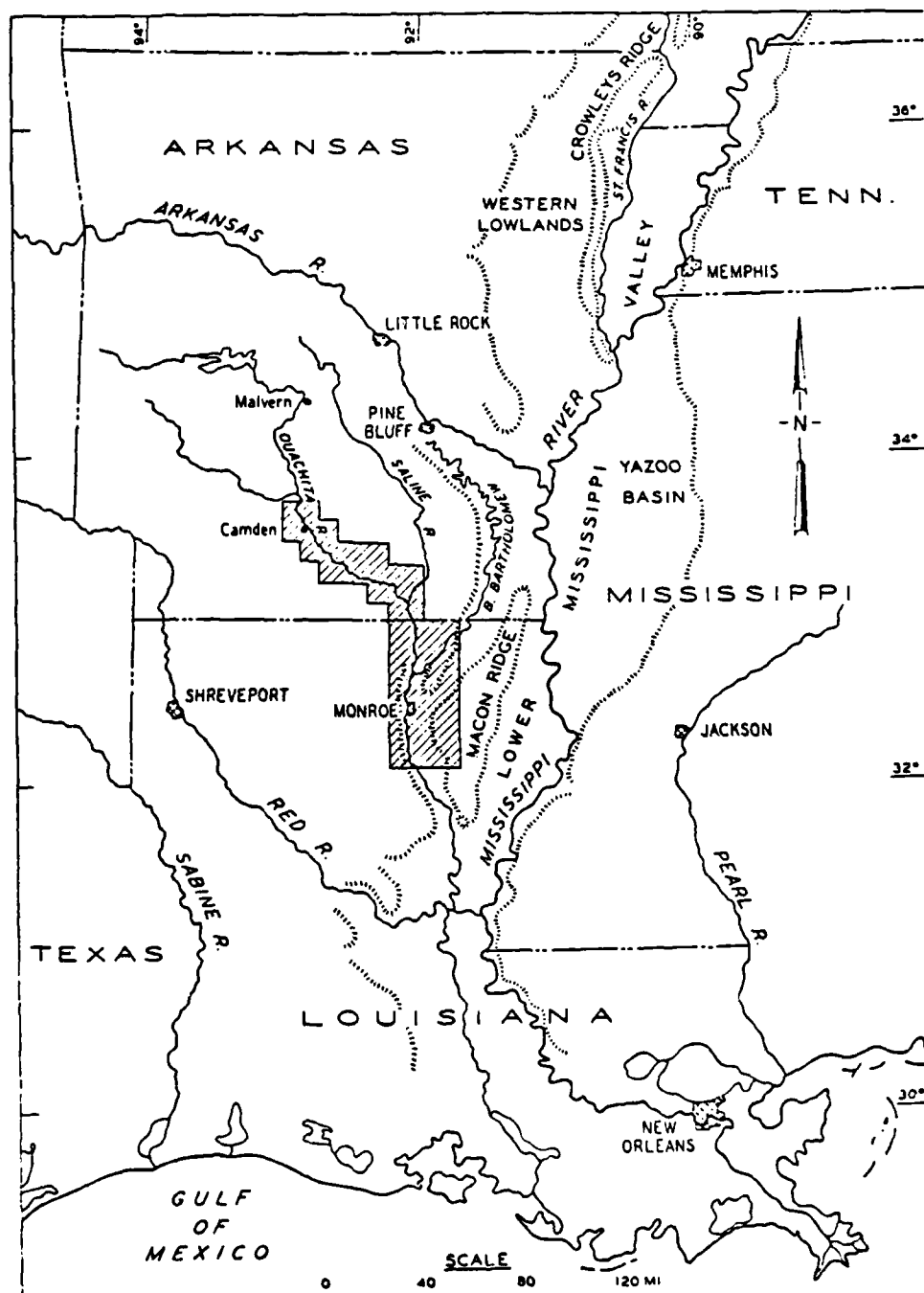
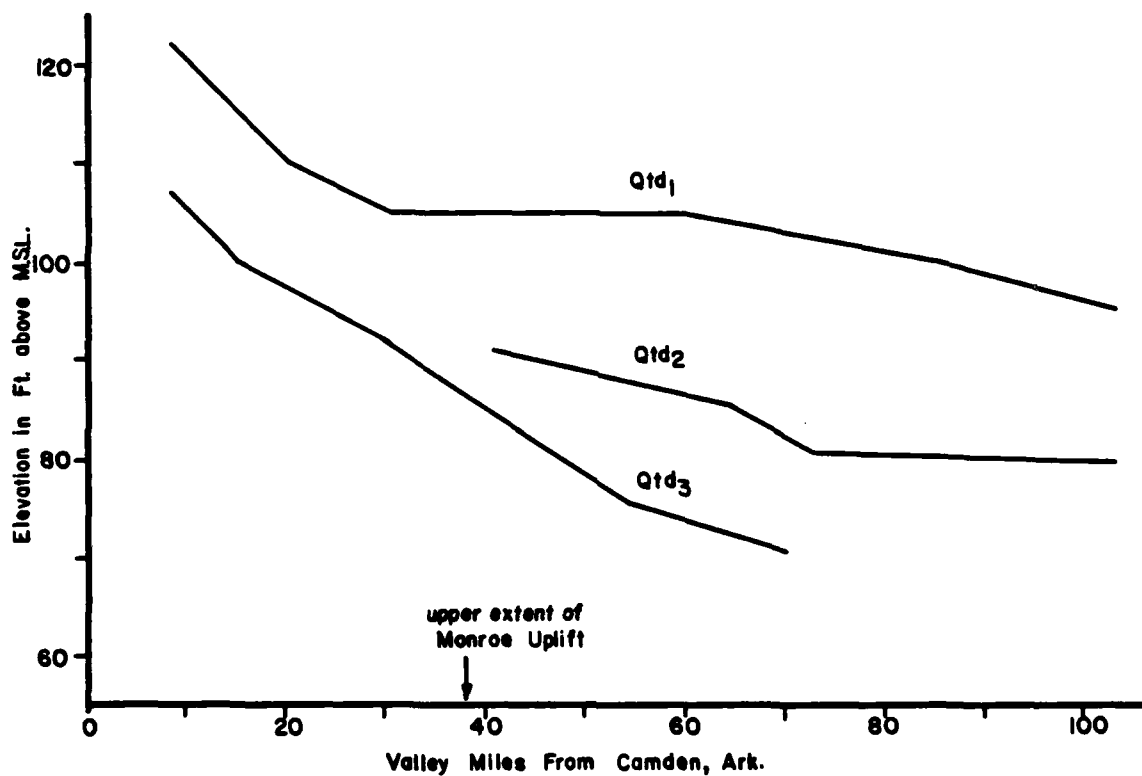


Figure 5.14 Deweyville Terrace profiles taken in hatched zone.
(from Saucier and Fleetwood, 1970)



Longitudinal Profiles of Deweyville terraces along the Ouachita River Valley.

Figure 5.15

Fleetwood (1970) (Fig. 5-16) and they identify these features as beach ridges which formed on the margins of an ancient lake, Lake Monroe, that was located in the Ouachita River valley during the Farmdalian Substage (Fig. 5-16). The maximum surface elevations of the ridge crests were examined using the surficial geology maps of Fleetwood (1969) and Saucier (1967) and they are shown with respect to valley location on Fig. 5-17. Several of these ridges are located within the Monroe Uplift area. The beach ridges circled on Fig. 5-16 have maximum elevations of between 110 and 130 feet, whereas all the other ridges to the south have maximum elevations of between 100 and 105 feet. If during mid-Wisconsin time, all of these ridges were formed along the margins of a lake then their crests should all be at approximately the same elevation, as stated by Saucier and Fleetwood (1970). However, the anomalously high crest elevations of the ridge segments, near the Louisiana-Arkansas border, suggest uplift. An alternative explanation for the discordant crest elevations is that the ridges to the north grew higher than those to the south or developed higher on the lake beach during the existence of the lake. However, the location of the anomalous ridges with respect to the Monroe Uplift suggests that uplift is the cause.

About 10,000 years ago, the Arkansas River became a meandering stream (Saucier and Fleetwood, 1970) and it has since created numerous meander belts across the Monroe uplift. Differences between the modern valley profiles along each of these channels and the proposed original profiles during formation of the meander belts should indicate deformation by earth movements.

At least three major channel courses and four minor courses have been identified using surficial maps of the area (Saucier, 1969).

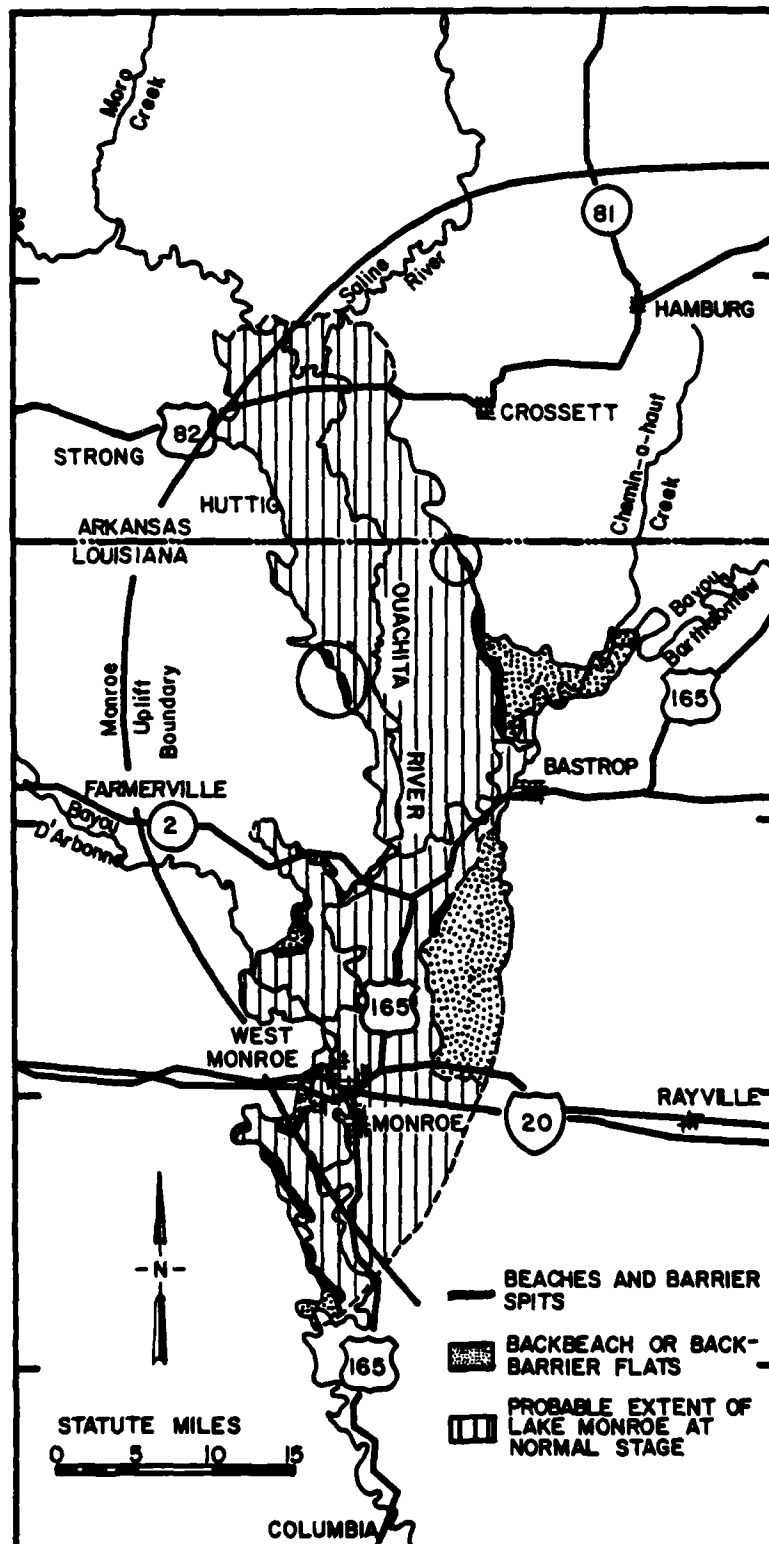


Figure 5.16 Location of Beach ridges along Ouachita River Valley. The circled ridges are 110-130 feet in elevation. (from Saucier and Fleetwood, 1970)

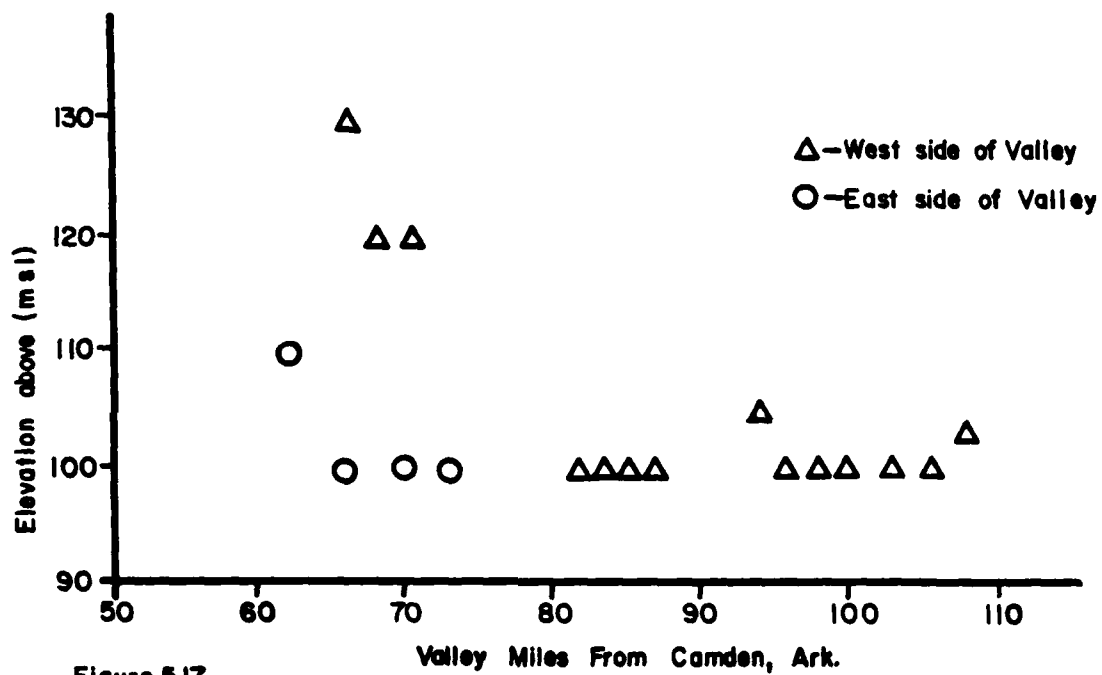


Figure 5.17

Elevations of Beach Ridges on Deweyville Terrace (Qtd.)
along the Ouachita River Valley.

Valley profiles of the backswamp lowlands along each of the courses were made (Fig. 5-18).

The backswamp lowland surfaces were chosen for valley profile construction instead of the natural levee crests because the levee heights were too variable and the levees were too discontinuous to produce a reliable and smooth valley profile. Although the backswamp elevations may be affected by areal differences in flood deposit accretion rates which would affect the shapes of the profiles to some degree, the gross irregularities in the levee profiles prevent recognition of post-depositional deformations.

It is evident from these profiles that a definite convexity occurs in the area of the Monroe Uplift. The original profiles should not have been convex but instead straight or concave up assuming that the Arkansas River was mature (Machida, 1960) and that its courses were developing during a slow progressive rise in base level. The margins of the convex reaches correspond fairly closely with those of the Monroe Uplift. The centers of the convex reaches, however, tend to be located in the south central portion of the Monroe Uplift area. The courses showing the highest degree of convexity occur on the western portion of the uplift. These courses are also the youngest being between 3500 and 1500 years old (Saucier and Fleetwood, 1970). These findings suggest that the uplift is still active with the axis and the most active zone located in the southern or possibly southwestern portion of the Monroe Uplift.

VALLEY PROFILES OF THE MODERN STREAMS

Assuming that the Monroe Uplift is still active today, the modern stream valley profiles should exhibit the effects of the uplift similar

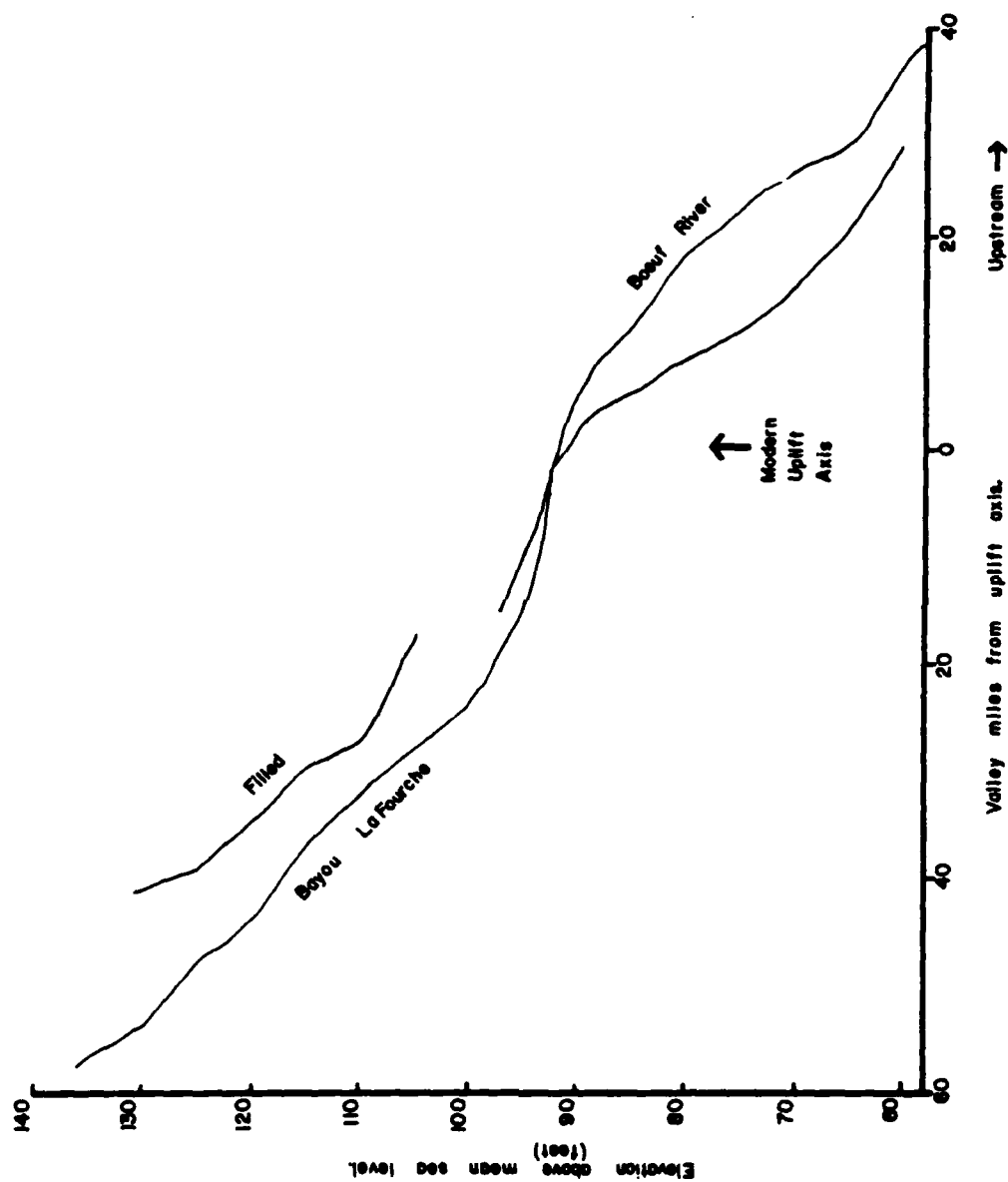


Figure 5.18

Valley profiles of old Arkansas channel courses in the Monroe Uplift area as they relate to the modern uplift axis.

to those shown by terrace and Arkansas River profiles.

In order to determine if this is the case the profiles of the Ouachita River, Bayou Bartholomew, Boeuf River, Big Colewa Creek, Bayou Macon and Deer Creek were analyzed. The profiles include the valley profile, the low water and thalweg profiles (which show elevation vs. channel length), and the projected thalweg profile (which shows thalweg elevation vs. valley length).

Of these various types of profiles, the valley profiles should most clearly exhibit deformation, which has been caused by recent earth movements. Unlike the profiles that use low water or thalweg elevations, the valley profiles are not as affected by changes in stream morphology such as sinuosity variability, aggradation or degradation within the channel, and changes in the width/depth ratio. The valley profiles are, however, affected by different rates of floodplain deposition along the valley. However, assuming that the streams in this study are mature, then according to Machida (1960), their valley profiles should be concave up if no valley floor deformation is occurring.

Valley profiles are plotted by measuring the elevation of the floodplain lowland, and the position of the measurement in the valley is plotted with regard to a straight line segment that follows the general trend of the channel course. Each valley profile shows an obvious zone of upward convexity and the profiles are arranged so that the convex zones of all the profiles are superimposed (Fig. 5-19). The Ouachita River valley profile is very irregular as compared to the other profiles. The marked upward convexity that occurs from 18 miles above the uplift axis to 25 miles below the axis probably represents a cross

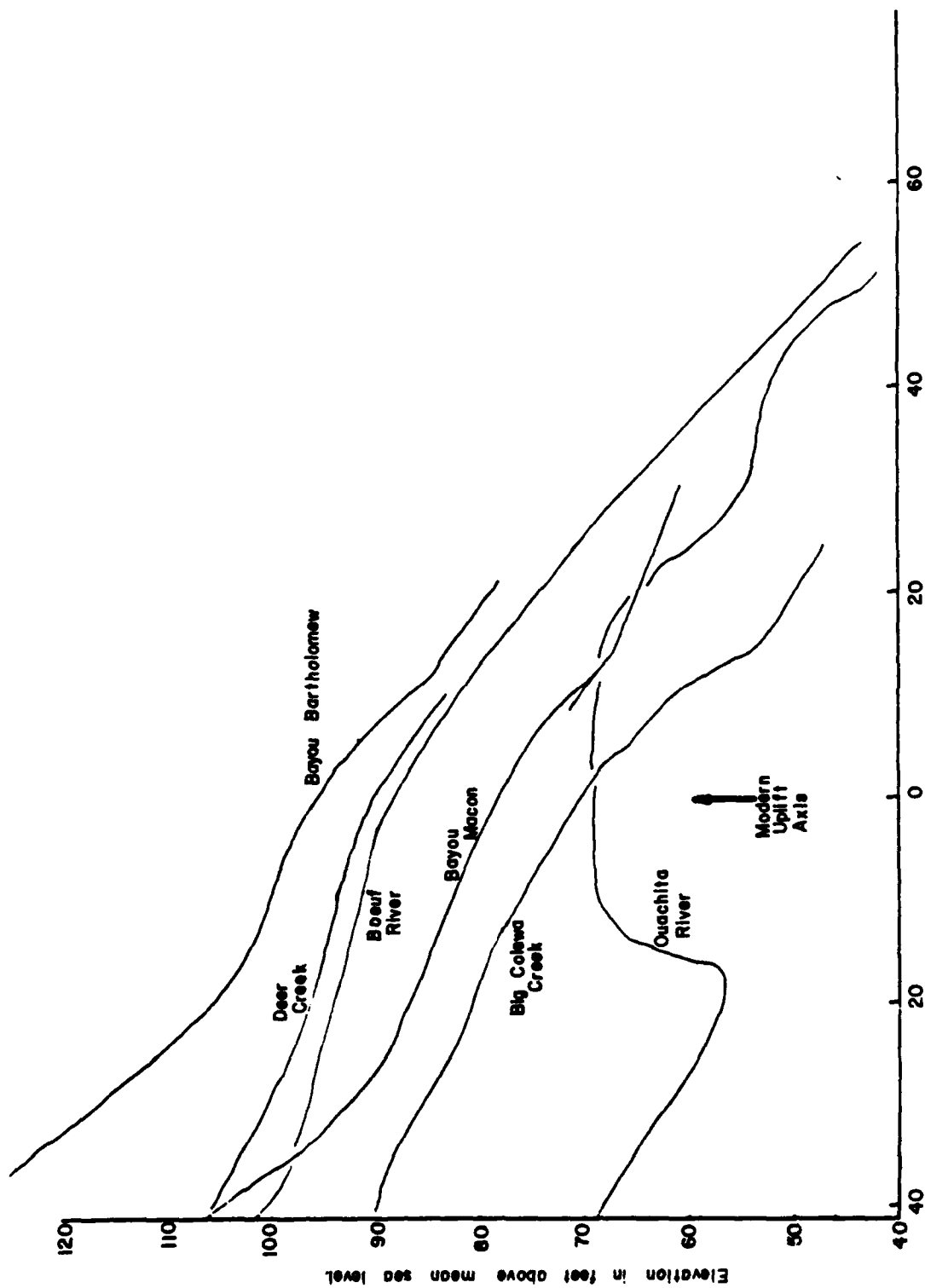


Figure 5.19
Valley profiles of six modern streams in the Monroe Uplift area in relation to the axis of the modern uplift.

section of the Arkansas alluvial fan which developed in the Ouachita River valley just north of Monroe, Louisiana. Therefore, the Ouachita River valley profile cannot be compared with the other five profiles.

It is clear from the valley profiles (excluding that of the Ouachita River) that an upward convexity of the modern floodplain surfaces occurs over the Monroe Uplift. Assuming that the center of the convexities on each profile represents the extent of the modern zone of uplift activity along each profile, then the limits and central axis of the modern uplift can be estimated.

THE PROPOSED ZONE OF THE MODERN UPLIFT

If the anomalous convexities and other deformities of the valley profiles of the Pleistocene fluvial terraces and Holocene streams in the Monroe uplift area are due to Quaternary uplift activity then these profile deformities can be used to determine the approximate boundaries and center or axis of recent uplift. Using this method, the proposed zone of recent uplift was mapped (Fig. 5-20). The proposed map shows the boundaries of recent activity as compared to the Tertiary boundaries of the Monroe Uplift and also the approximate axis of the recent domal uplift activity. The proposed axis runs through the centers of the convexities of the valley profiles and the upper and lower limits of these convexities on the profiles define the proposed boundaries of recent uplift.

In the eastern portion of the Monroe Uplift, the center of uplift activity appears to trend east-west just north of the southern boundary. In the central portion, the axis follows a north-south trend, and then in the north-west portion of the uplift area, it trends to the west through the vicinity of Felsenthal, Arkansas. The Monroe Fault is

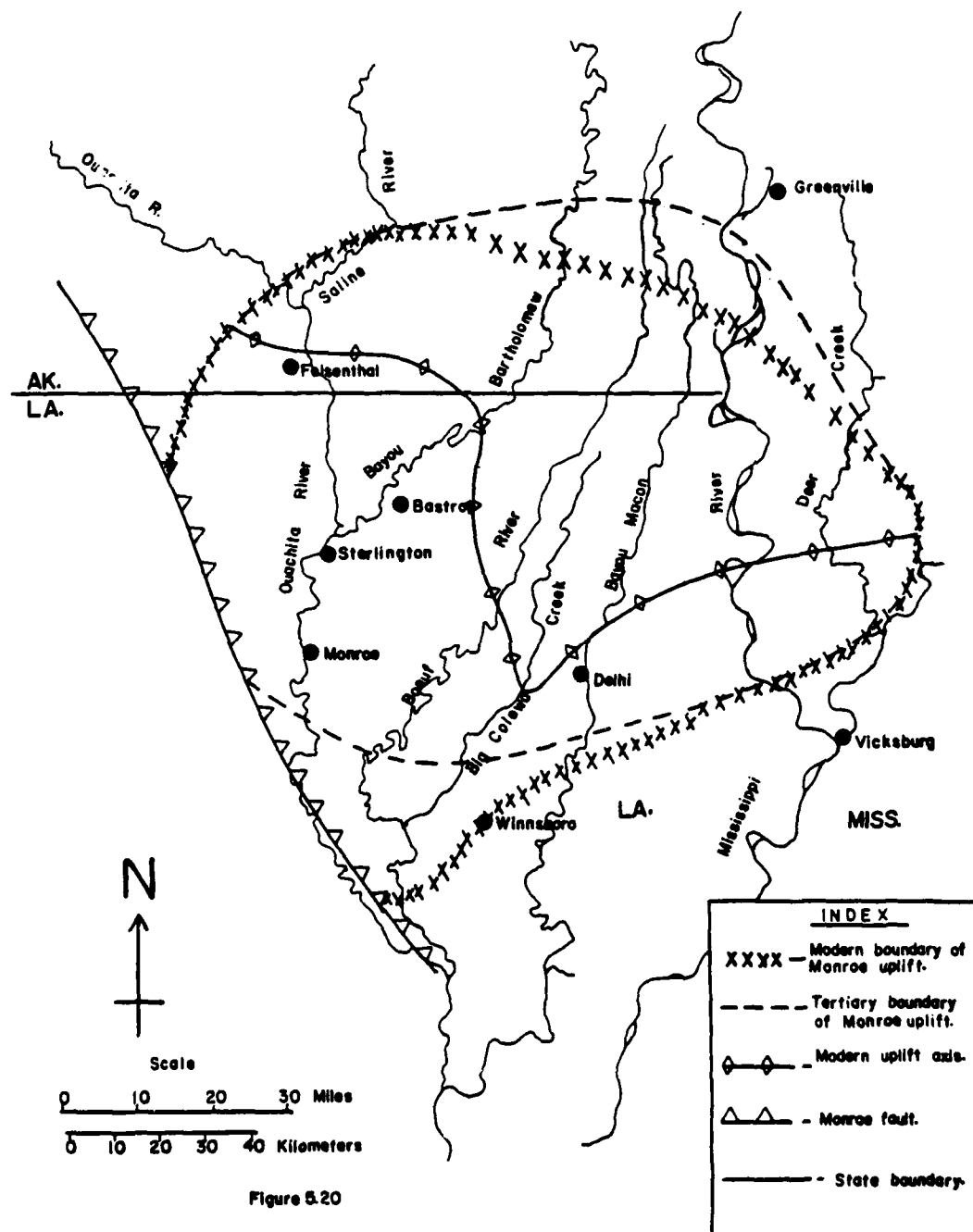


Figure 5.20

positioned along the south-western boundary of the recent uplift area and extends further south-east to Sicily Island.

The present location of the Monroe Fault on the surface is based on the locations of the southeast trending valleys of the Ouachita and Boeuf Rivers south of Monroe, Louisiana, the location of the southeast trending portion of the Highland scarp, and also the locations of the parallel southeast trending valleys dissecting the Highlands.

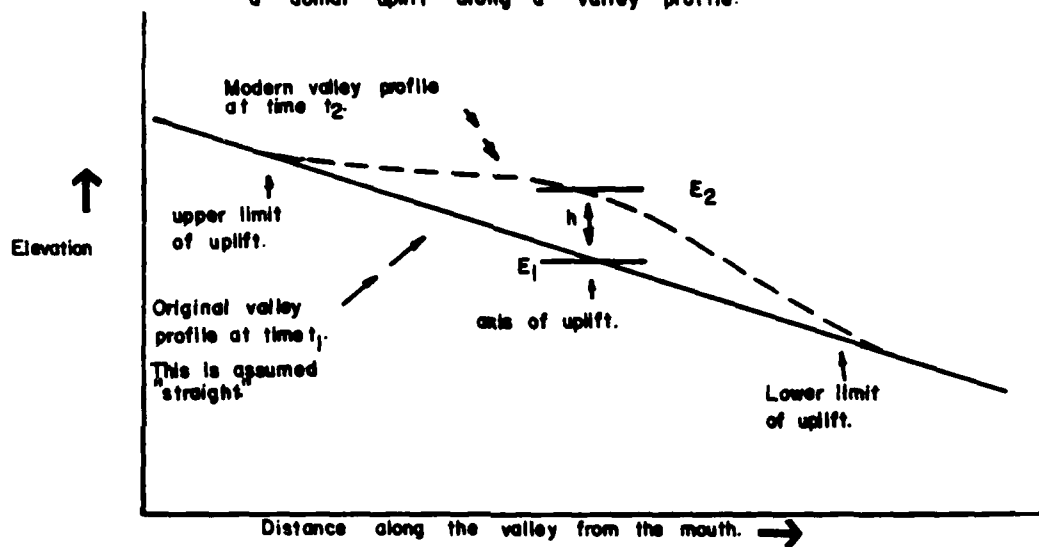
MODERN RATES OF UPLIFT

Comparing the amounts of vertical deformation of the terraces and floodplains in the Monroe Uplift area to their ages of formation provides a means of estimating contemporary rates of uplift in this area.

The amount of uplift can be determined as the maximum elevation difference between the actual convex profile and the original profile which crosses through and under the convexity in the modern profile. If the original profile is assumed to be fairly smooth then the maximum amount of uplift can be estimated (Fig. 5-21). By this method, the convexity in the oldest and highest Macon Ridge terrace profile (Qtb_1) indicates that about 12 feet of vertical deformation occurred at the uplift axis since this terrace was formed. Saucier (1970) estimates the age of this terrace to be 33,000 years old. Therefore, the maximum rate of uplift is estimated at 1.1 mm per 100 years over the last 33,000 years.

It is assumed that the floodplains bordering the modern streams were originally smooth with fairly straight (to concave) profiles. Elevation differences between the actual convex profile of each of the streams and a straight line profile running beneath the convexity

Method for estimation of vertical deformation due to a domal uplift along a valley profile.



$$h = E_2 - E_1 \quad \text{where}$$

h = maximum amount of vertical uplift of the valley surface from t_1 to t_2 .

Figure 5.21

E_2 = modern elevation at the axis of the uplift (at t_2)

E_1 = original elevation at the axis of the uplift (at t_1)

$$\text{Rate of uplift} = h / (t_2 - t_1)$$

indicates that between 5 and 10 feet of vertical deformation along the uplift axis has occurred.

Because the Arkansas River carried a very large sediment load compared to the modern streams which are using its abandoned courses, the surrounding floodplains were probably most extensively developed during the Arkansas River's occupancy of the Monroe Uplift area. Since the time when the Arkansas River changed its course at Pine Bluff, Arkansas, the floodplains surrounding its old courses over the Monroe Uplift probably have not aggraded appreciably. Saucier and Fleetwood (1970) estimate that only a few feet of floodplain deposits have accumulated since the removal of the Arkansas River source of sediment. Therefore, the dates of major course changes of the Arkansas River are used to estimate the ages of the surrounding floodplains in the Monroe Uplift area.

Saucier and Fleetwood (1970) estimate that the Arkansas River course used presently by Bayou Macon was abandoned 3000 to 3500 years ago. Because the Bayou Macon profile shows a vertical deformation of 5 feet at the uplift axis, a rate of uplift of about 44 mm per 100 years over the last 3500 years is inferred for the central portion of the uplift.

The modern Boeuf River profile shows 10 feet of vertical offset at the uplift axis compared to a straight line profile. Since the Old Arkansas course following portions of the Boeuf River is estimated to be 3000 years old, a rate of 100 mm per 100 years or 1 mm/year is inferred for the Boeuf River region of the Monroe Uplift.

The Arkansas River abandoned the Bayou Bartholomew course approximately 1000 to 1500 years ago (Saucier, 1970) and the Bayou

Bartholomew profile shows a vertical offset of 4.5 feet at the uplift axis. Therefore, a rate of vertical deformation of between 0.9 and 1.4 mm/year during the last 15,000 years, is estimated for the axis at the point crossing Bayou Bartholomew.

The calculated uplift rates vary between 0.01 and 1.4 mm/year for the modern Monroe Uplift. In general, the more recent periods of deformation show the highest rates of uplift. This trend may imply that the activity of the Monroe Uplift has been increasing over the last 30,000 years or that uplift is episodic.

EVIDENCE OF EFFECTS ON STREAM MORPHOLOGY

If the Monroe Uplift has been active during the Pleistocene and Holocene epochs, then evidence of particular responses of the past and present streams to the uplift activity is expected. Therefore, the fluvial morphology of the area was examined to determine whether such evidence of stream response actually exists in the area.

Lateral or transverse tilting of a valley can cause a river to shift in the downtilt direction (Fig. 3-7). Based on this concept, if the Monroe Uplift has been recently active then particular patterns of the ancient and modern drainage systems should have been produced in response to the uplift. The existing drainage system as well as the pattern and chronology of the ancient Arkansas River channels do conform with the expected responses to surface deformation.

Immediately preceding the early Wisconsin regression, the Arkansas River aggraded its valley to the level of the highest Macon Ridge terrace. Then it shifted to the west side of the valley, as it downcut (Saucier and Fleetwood, 1970), where it actively attacked the western valley margin. This westward shift could have been in response to a

tilting of the valley surface to the west or away from the center of uplift which was located further to the east.

During the last 10,000 years the modern meandering Arkansas River has created many different channel courses in the Monroe Uplift area due to a series of major avulsions (Fig. 5-22). Analysis of the relative ages of the channel courses show that the net movement of the channels has been to the west. A series of course changes from east to west through time may be due to westward tilting of the surface from Macon Ridge to the western boundary of the valley.

The modern drainage pattern in the Uplift area also appears to reflect recent effects of landsurface tilting (Fig. 5-23). Bayou Bartholomew flows due south in the area north of the Uplift, and then immediately upon entering the uplift area it begins a southwesterly course. Boeuf River and Big Colewa Creek also flow to the southwest across the Monroe uplift. Ouachita River flows eastward until it encounters the Uplift, and then it abruptly changes direction to the south-southwest. Likewise the Saline River makes an abrupt direction change to the southwest before entering the Ouachita River. These trends all could be caused by a westward tilting of the floodplain surface within the western half of the Uplift. The westward trend should not normally be expected since the Mississippi Valley trend is to the southeast.

Upstream of the axis of Uplift there should be a decrease of the valley-floor slope due to backtilting. Downstream of the axis of uplift there should be an increase of valley-floor slope. A stream crossing the Uplift should be affected by such slope changes by a reduction of

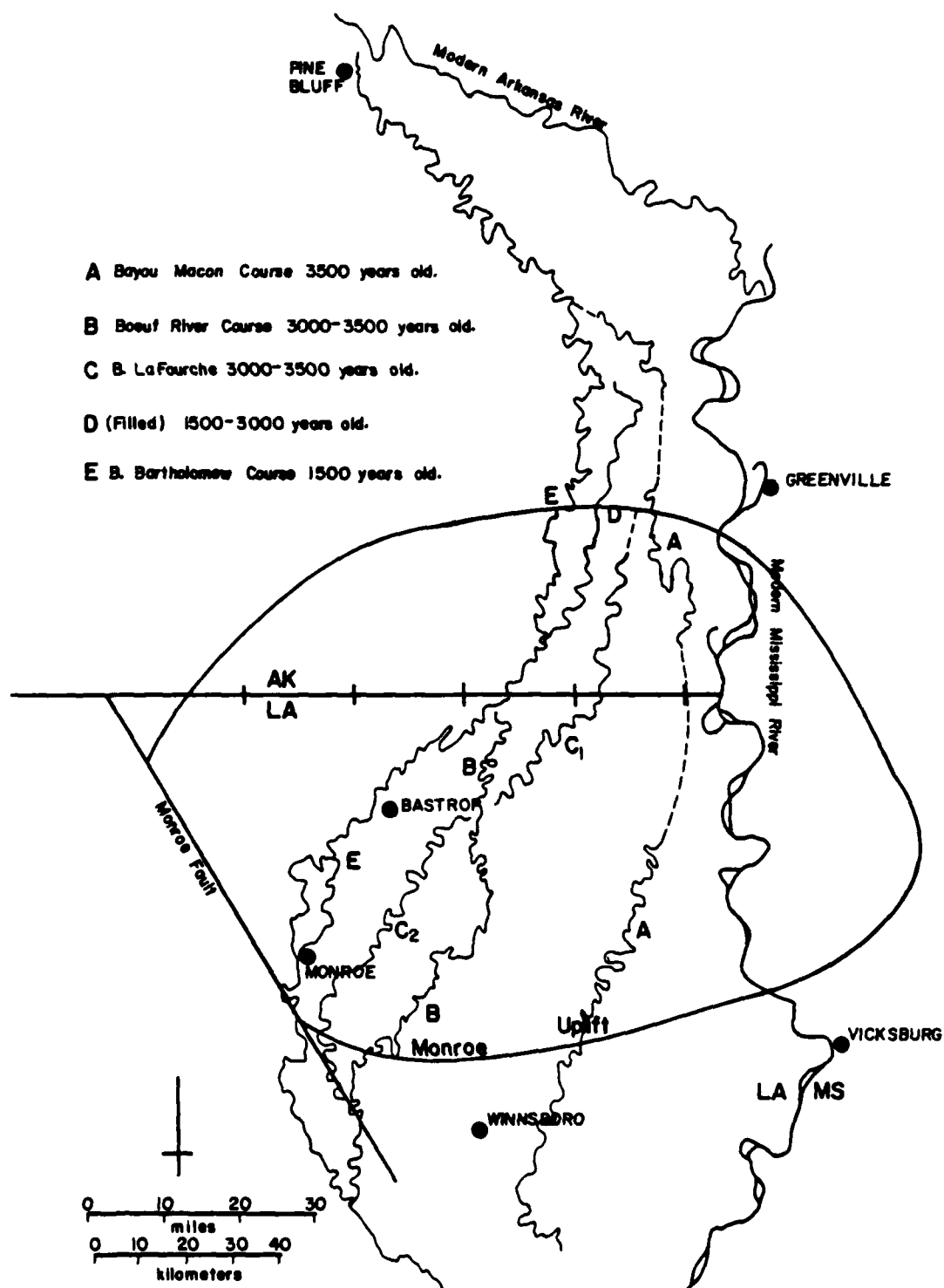


Figure 5.22

Old Arkansas channel courses
in the Monroe Uplift area.

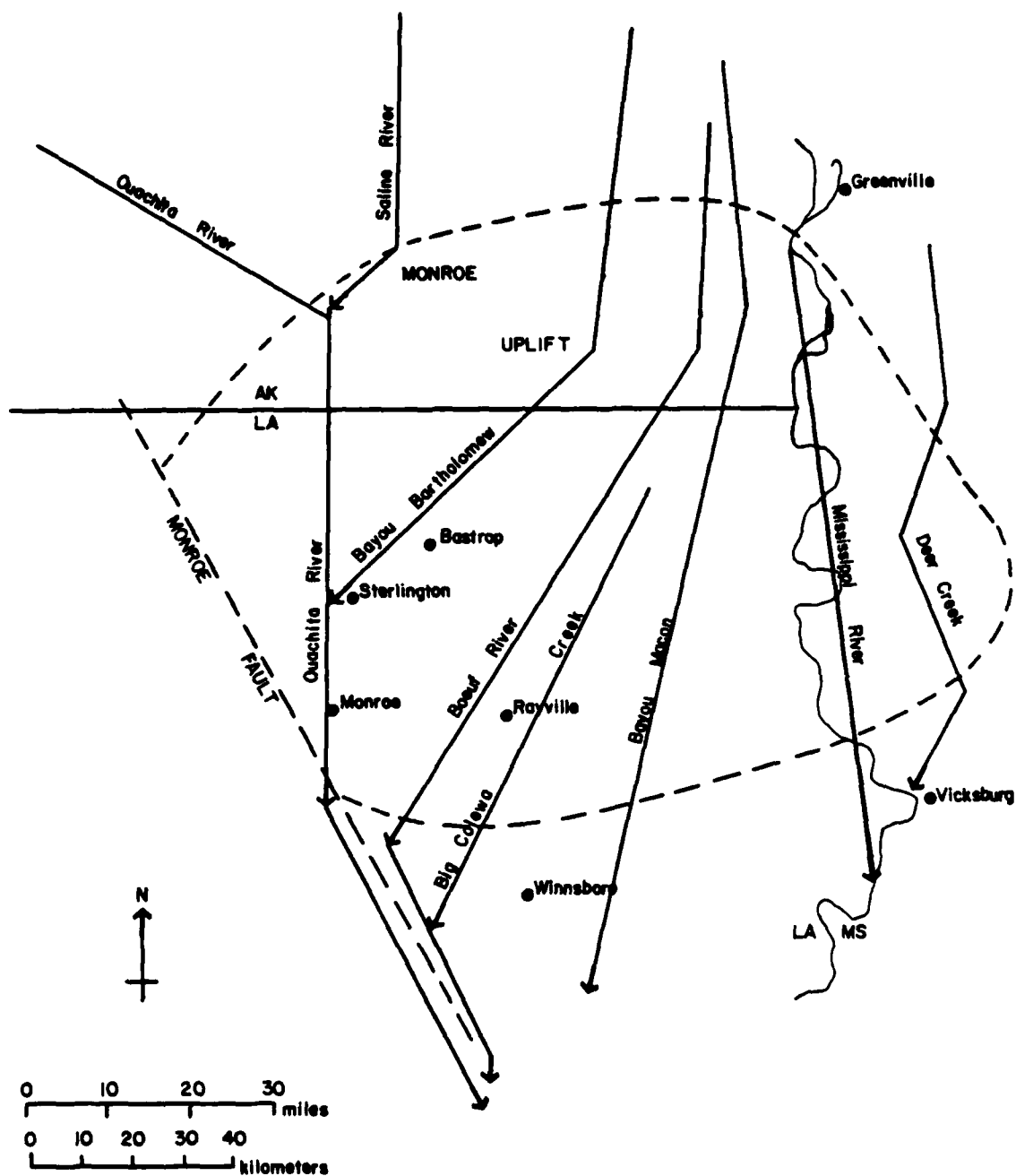


Figure 5.23 Schematic diagram of channel patterns as they relate to the Monroe Uplift.

competence and stream power in the area upslope of the axis and by an increase of competence and stream power downslope of the axis. The stream should respond to such changes.

Decreased slope above the axis should result in aggradation, assuming that sediment load from upstream does not decrease (Schumm, 1972). Also, decreased valley slope can cause various stream-channel pattern changes such as a change from a branchwork pattern (Fig. 2-4) to a network pattern (Russell, 1939) and a change from a single meandering channel to an anastomosing pattern (Fig. 3.3).

Evidence exists that these responses have indeed occurred immediately upslope of the proposed center of the Monroe Uplift. The confluence of the Saline and Ouachita Rivers exhibits a complex network pattern with a multitude of diverging and converging stream channels. This pattern occurs on both the major and minor streams in this area. The pattern probably developed, as the streams aggraded in response to slope reduction, and then broke out of their channels in numerous places to create a distributary pattern. The backswamp lowlands in this area are covered by well-developed swamps and numerous lakes indicating poor drainage. This backswamp drowning may also be in response to a reduced gradient of the floodplain surface due to backtilting.

According to Saucier (personal communication), the Saline River carries a coarse sediment load that should have an effect on the morphology of the confluence of the Saline and Ouachita Rivers. Rapid alluviation at the confluence from the Saline sediment load could cause damming of the Ouachita River above the confluence. However, the large-scale nature of the anastomosing pattern, which occurs both above and below the confluence, as well as the absence of an alluvial cone at

the mouth of the Saline River suggest that the sediment load does not provide an adequate explanation for the existing morphology.

Thicker than normal alluvial deposits are evident along the Western Highlands north of the Monroe Uplift, as compared to downstream areas (Saucier, 1967). These were deposited by extensive alluviation of the streams entering from the Highlands into the Mississippi Valley. The Cutoff Creek basin, shown on the Collins, Arkansas, topographic sheet, once had a branch work pattern (Fig. 2-4). Now the stream gradients are reduced to nearly zero and a swamp network has developed. In contrast, the valleys within the Highlands scarp south of the Monroe Uplift center are not as extensively alluviated and do not show clear network patterns.

MORPHOLOGY OF THE MODERN CHANNELS IN RELATION TO THE UPLIFT

Using the boundaries and axis of the Monroe Uplift as defined (Fig. 5-20), relationships between current vertical deformation of the earth and the morphology of the modern stream channels crossing the uplift area can be examined. A pattern of morphologic changes should be detected for a stream which is responding to contemporary uplift within its valley.

SINUOSITY

Adams (1980) found that a correlation exists between the downstream change in sinuosity of several rivers in the mid-continent and tilt rates measured along their valleys. The sinuosity of a stream over any particular reach is defined as the ratio between the channel thalweg length and the corresponding valley length. He suggests that the rivers are responding to the valley slope changes by changing their sinuosity in order to maintain equilibrium slope. Since downstream sinuosity changes may record valley deformation, as suggested by Adams, the sinuosity values of six of the modern streams crossing the Monroe Uplift

were measured (Fig. 5-24).

The sinuosity values were calculated for two, five, and ten mile segments of the stream valleys in order to produce representative values of sinuosity for each valley. For each plot the location of the proposed uplift axis is shown. The axis locations were determined from the convexities in the valley profiles of the streams. From these plots it is apparent that in every case the downvalley side of the uplift axis has higher values of sinuosity than immediately upvalley of the axis. It also appears that a rather sudden increase in sinuosity occurs in the vicinity of the uplift axis. Therefore, the low sinuosity values correspond closely to the low valley slopes found upvalley of the uplift axis. Likewise, higher sinuosity values correspond to the higher valley slopes found downvalley of the uplift axis. These findings imply that the streams are adjusting to the uplift in such a way as to decrease their sinuosity along those parts of the valley which show a reduced slope (i.e., upvalley of a domal uplift) and to increase their sinuosity along the valley segments showing an increase in slope (downvalley of a domal uplift) (See Table 3-1).

These modern streams, except the Big Colewa Creek, incorporate at least in part, the highly sinuous courses of the Arkansas River. The patterns of occupation of the old course conform to the location of the proposed uplift activity. Upstream of the uplift axis, the Boeuf River and Bayou Macon have relatively straight courses and do not incorporate an old Arkansas course. Downstream of the axis they do incorporate old courses. Comparison of the meander wavelengths and radii of curvature of the modern channels with those of the old Arkansas courses indicates that the modern streams have reworked the older channels which they

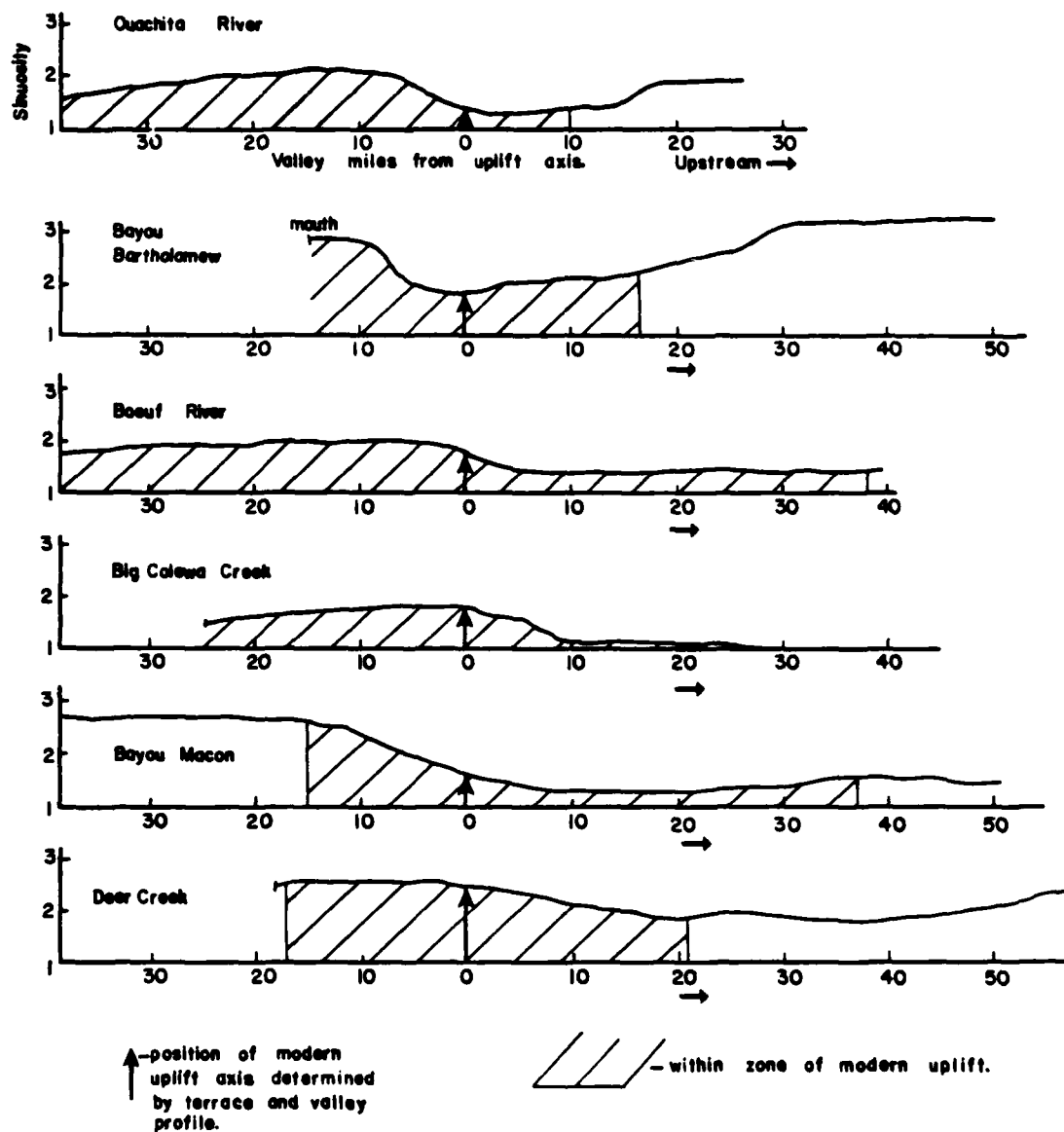


Figure 5.24

Variations of sinuosity along valleys of six Monroe Uplift streams. The zone and position of axis of uplift were determined from valley profiles. (Fig. 5.19)

occupy. In all cases, in the stretches of occupation downstream of the axis, the sinuosity of the modern channels is higher than that of the paleo-channels. In contrast, upstream of the axis, the channels of Boeuf River and Bayou Macon both cross old aggraded Arkansas courses several times without following the sinuous paths.

These observations imply that in the uplift area above the axis, the modern streams have too low a gradient to maintain the highly sinuous course of the Arkansas River channels, and instead they have aggraded these channels and created their own straighter and more efficient channels. Downstream of the uplift axis the modern channels have been able to transport sediment through the old sinuous courses, as well as to rework the older courses to produce courses with even higher sinuosity. Upstream of the axis, the valley may have backtilted to such an extent that the energy of the modern channels is significantly reduced. In response to such backtilting, the streams could have aggraded their channels and formed relatively straight channels. Downvalley of the uplift axis, downtilting of the valley could have increased the gradient of the modern streams, allowing them to incorporate and rework the old courses to produce highly sinuous courses. Field evidence suggests that the Boeuf River has also degraded the old Arkansas River channel downstream of the uplift axis. In the area just north of Rayville, Louisiana, at least two levels of paired benches occur along the channel within the steep narrow valley which defines the old Arkansas course. Adjacent oxbows once formed by the Arkansas River are at least 20 feet higher in elevation than the present Boeuf River channel. In this area the present banks of the Boeuf River are about 40 feet high.

PROJECTED CHANNEL PROFILES

In order to examine variability of the channel bed elevation and also changes in the bank height along the valley of the streams which cross the Monroe Uplift, channel elevations were plotted in relation to the valley distance along Boeuf River and Big Colewa Creek. These plots of channel elevation to valley distance are called the projected channel profiles of the streams. The projected channel profile is not affected by changes in sinuosity because the thalweg elevation (or low water elevation) at a given location is plotted with reference to valley distance. The projected channel profile is useful because it clearly indicates the degree of aggradation or degradation which has occurred in the channel in relation to distance along the valley. Thus, using the projected channel profile, degradation or aggradation of the bed can be related to location within the uplift area.

Figure 5-25 shows the projected channel profiles of Boeuf River and the Big Colewa Creek in relation to their valley profiles and to their position on the Monroe Uplift. Several observations can be made. First, the difference in the elevation of the projected channel profile and that of the valley profile at a given location represents the depth of the channel below the valley surface at that location. This elevation difference will be called the average bank height. Those reaches with large differences in elevation between the valley surface and channel are where the channel has downcut or degraded (assuming fairly constant alluviation of the floodplain along the valley). Likewise, those reaches in which average bank height is small indicates that the channel has not degraded or has, in fact, aggraded. The projected channel profiles are not parallel to the valley profiles

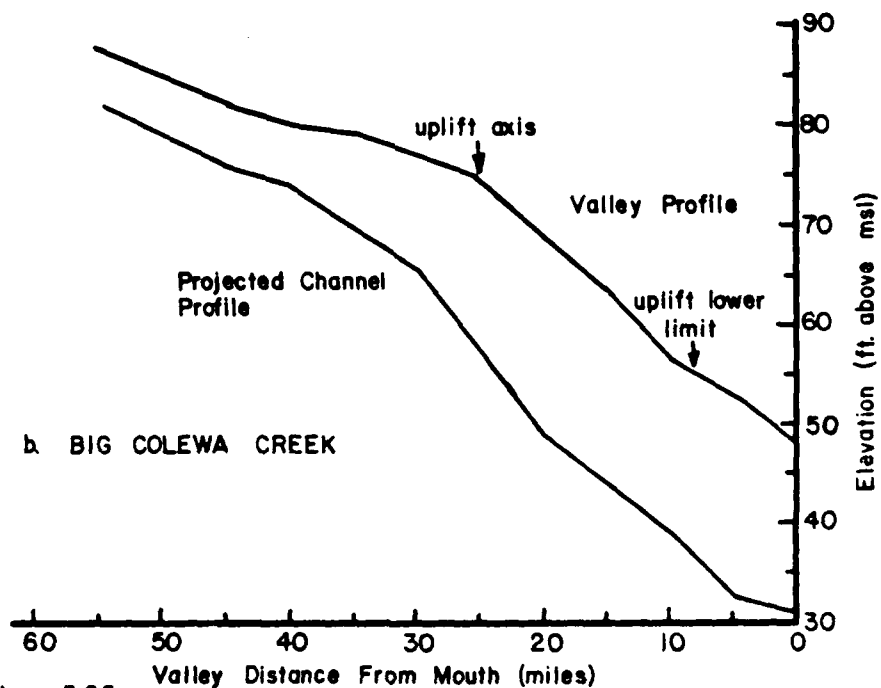
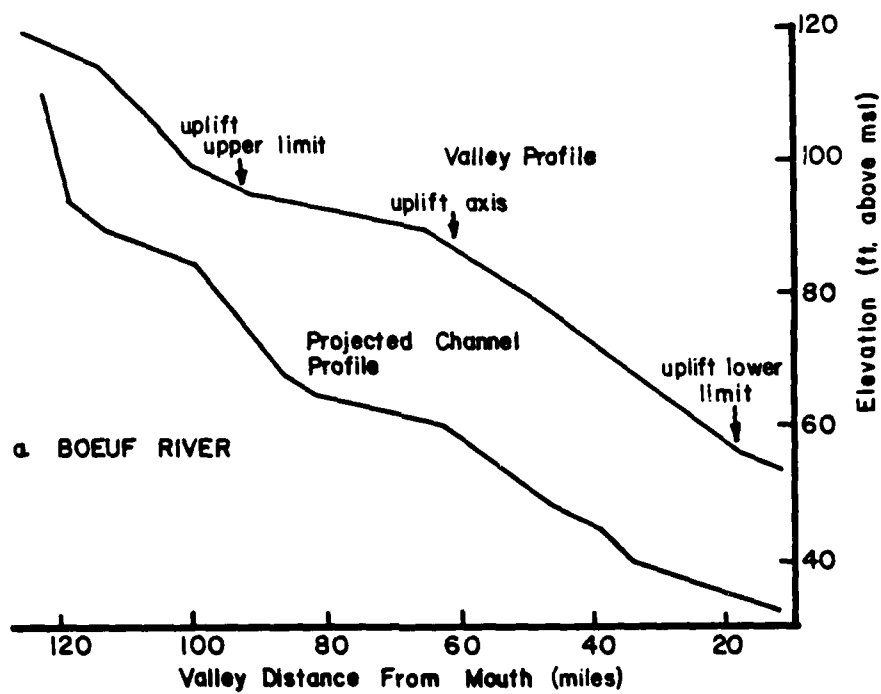


Figure 5.25 Valley profiles and projected channel profiles of Boeuf River and Big Colewa Creek.

indicating that varying amounts of degradation or aggradation have occurred along the channel (Fig. 5-25). The projected channel profiles, in both cases, have convex sections that occur within the boundaries of the present uplift (Fig. 5-20). However, the apex of each of the convexities on the projected channel profiles lies upstream of the uplift axis. At the axis of the uplift and in the downvalley zone of the uplift, the average bank heights are, in general, very high (35 to 40 feet for Boeuf River and 20 feet for Big Colewa Creek). Upstream of the uplift axis, bank heights decrease (as low as 12 feet along Big Colewa Creek). These observations indicate that entrenchment may be occurring at and below the uplift axis. Above the uplift entrenchment has not occurred.

The steepest segments of the projected thalweg profile should indicate where the most active degradation is occurring, and it is possible that these stretches of active degradation may be migrating upstream through the uplift axis (Fig. 3-4).

Degradation of the channel may be a response to active uplift and tilting of the valley. Entrenchment in the downvalley part of an uplift, where downtilting is increasing the slope, would tend to reduce the channel slope in the downvalley stretch, and tend to increase the channel slope upstream of the entrenched stretch.

FURTHER ANALYSIS OF BIG COLEWA CREEK

Big Colewa Creek is the only major channel across the Monroe Uplift. The old Arkansas River courses tend to modify the channels which incorporate them by affecting their sinuosity, channel slopes, and levee heights. Therefore, the Big Colewa Creek channel should best represent the effects of the Monroe Uplift on stream morphology.

The Big Colewa Creek channel can be broken into three zones of activity along its valley (Fig. 5-26a). The lower zone, from the mouth to valley mile 20, has a rather constant high average bank height of about 19 feet. The middle zone, from valley mile 20 to 40 shows a clear upstream decrease in the average bank height from 20 feet to 6 feet. The upper zone, above valley mile 40, shows a constant low average bank height of 6 feet. The lower zone may be where entrenchment has already occurred. The middle zone may be where entrenchment is still in progress and the upper zone where entrenchment has not yet occurred.

Figure 5-26b shows changes in valley slope and channel thalweg slope with valley distance. The valley slope remains large (about 1.2 ft/mile) from the mouth to valley mile 25 and then it suddenly drops to a low constant value of about 0.5 ft/mile. The break in slope, at valley mile 25, defines the proposed location of the uplift axis. The thalweg slope remains constant at a high value of 1.0 ft/mile from above the mouth to valley mile 35 and, likewise, it suddenly decreases to a low value of 0.5 ft/mile.

The sinuosity is approximately 1.2 in the upstream stretch of Big Colewa Creek (Fig. 5-26c). At valley mile 30, the sinuosity suddenly increases to above 1.7 in a downstream direction, and then gradually reduces to a value of 1.5 at the mouth.

Comparing average bank height to valley slope, the middle zone of active entrenchment is centered on the axis of the uplift. This zone of active entrenchment could also be in the process of migrating upstream through the center of the uplift because the initial stage of the bank height increase (or entrenchment) occurs upstream of the uplift axis (mile 25) where, in fact, the valley slope is very low.

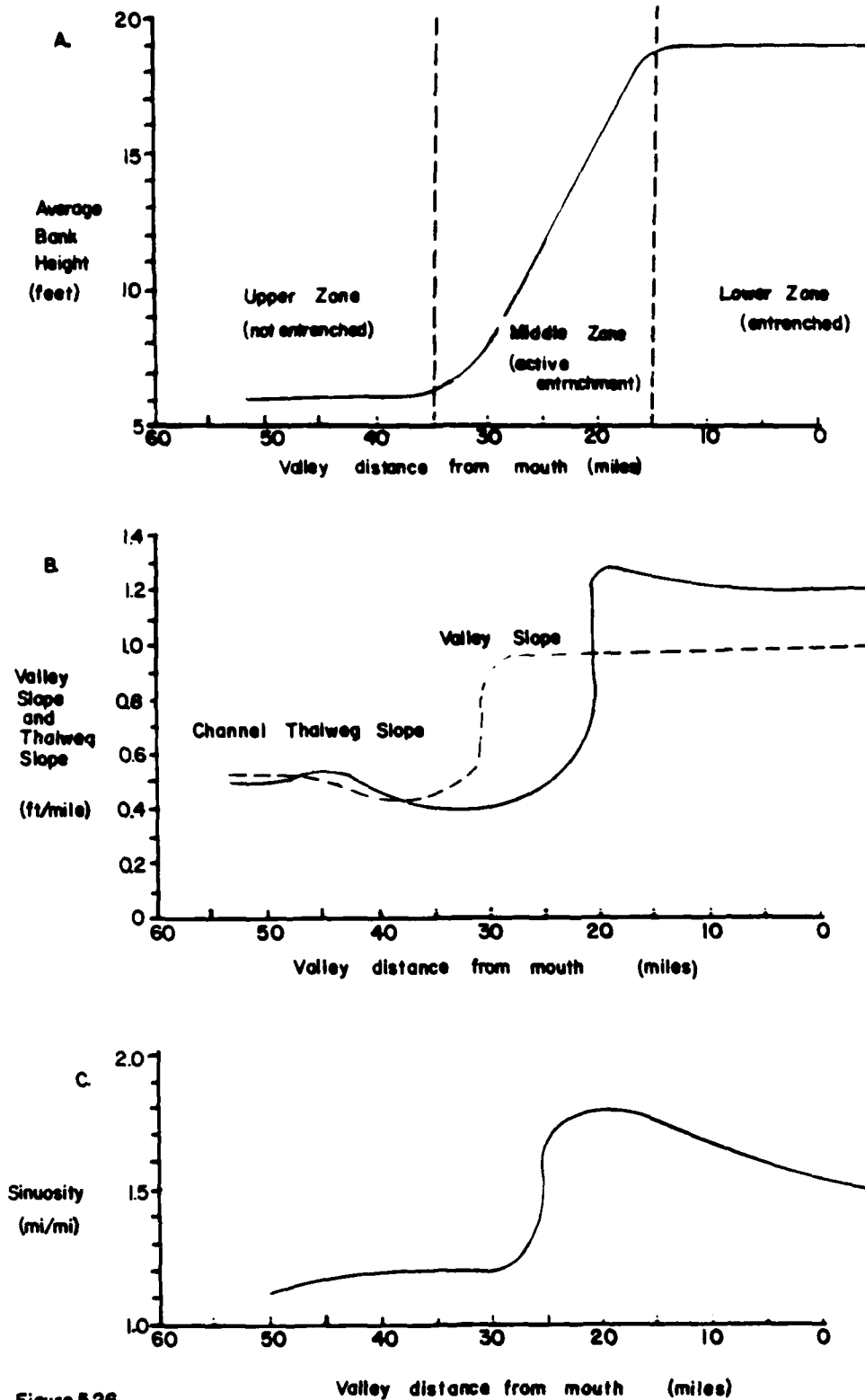


Figure 5.26

Variations of bank height, gradient and sinuosity, Big Colewa Creek.

A Corps of Engineer profile (Vicksburg District, 1940) shows that an anomalously deep thalweg exists between valley miles 22 and 26 (within the zone of active entrenchment). The average thalweg along this stretch of channel is five to ten feet deeper than that immediately upstream or downstream. This local scouring may represent a knickpoint in the channel bed and the shallowing of the thalweg downstream may be due to local aggradation of sediment which was mobilized by upstream knickpoint migration. It is unlikely that variation in bed and banks is sufficient to explain this local scouring.

The graph of average bank height to valley slope (Figure 5-27) shows that those points from downstream of the uplift axis and those points upstream of the uplift axis are separated into two distinct groups. Downstream of the axis the channel has high average bank heights and high valley slopes. Upstream of the axis the channel has low average bank heights and low valley slopes. As a whole the graph shows a clear separation between average bank height and valley slope. This suggests that the degree of entrenchment of the channel is related to the amount of downtilting (or increase in slope) that has occurred due to the uplift. If the relation is, indeed, one of cause-and-effect then it shows that entrenchment of the channel through the uplift axis may be a direct response to the valley tilting.

Comparing channel slope to average bank height, it appears that the channel gradient does not remain constant, as assumed by Adams (1980), but it rapidly increases through the zone of active entrenchment and remains high downstream where entrenchment is complete. The mobilization of channel bed sediments due to degradation may require a higher channel slope to effectively carry the increased sediment load.

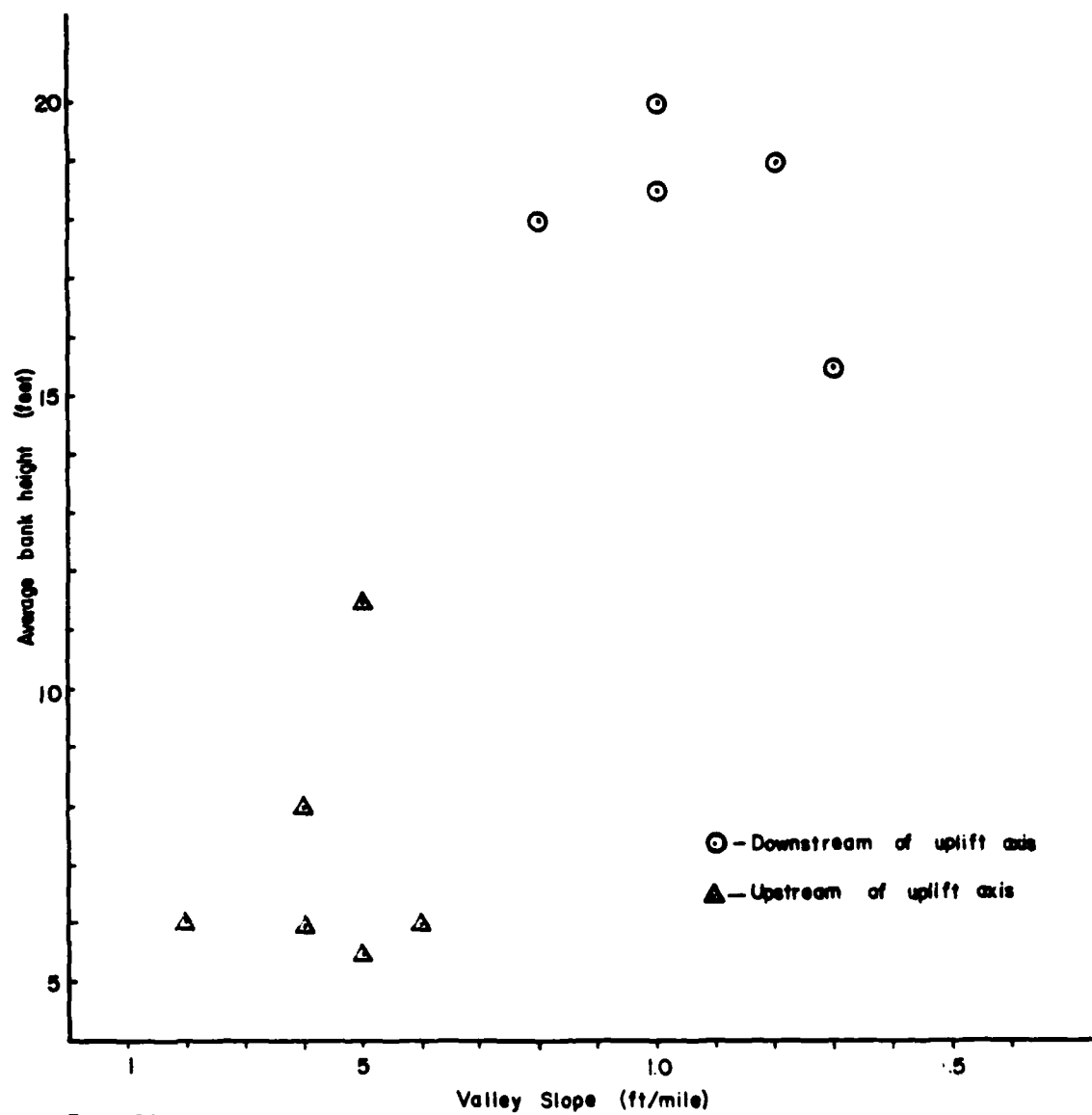


Figure 5.27

Difference between bank height and valley slope above and below uplift axis.

There will be a much lower sediment load upstream of a zone of entrenchment, and therefore, lower channel slope is required to transport the sediment.

Comparing sinuosity to average bank height, the sinuosity of the channel does not increase in a downstream direction until a certain level of entrenchment has occurred. Where the average bank height exceeds 12 feet, the sinuosity rapidly increases and remains at a high value. This relation may indicate that there is a threshold bank height which, if exceeded, results in bank instability and collapse by toppling or slumping. The increased thalweg slope through the stretch should accelerate the bank failure and widening by providing the stream with enough energy to remove the loosened bank material. This widening is coupled with increased lateral stream migration, point bar development, and an increase in sinuosity through the steepened stretch.

Downstream of the zone of active entrenchment the sinuosity decreases slightly. This decrease in sinuosity may be due to an increased sediment load from upstream active entrenchment and bank erosion. This may cause the channel to increase in slope and thus to straighten its course in order to effectively transport the increased sediment load.

CONCLUSIONS

Numerous relations between the morphology of the streams and terraces of the Monroe Uplift area and the underlying Cretaceous and Tertiary structures of the Monroe Uplift suggest quite strongly that the area is still active. Assuming that the Monroe Uplift is still active, then the patterns of changes which have occurred on the stream systems in the area and the present morphology of the streams provide an

interpretation of how streams respond to active domal uplift within their valleys.

The channels of six streams which cross the Monroe Uplift area were examined for possible effects of recent uplift activity. From this morphological examination, a pattern of channel responses to domal uplift activity is proposed.

The initial response of an alluvial stream to an upwarping of its valley may be entrenchment. It appears that the entrenchment begins below the uplift axis in the zone where the channel is becoming steepened through time. The zone of entrenchment may then progress upstream through the axis of the uplift. This zone of active entrenchment creates a reach with a very steep projected thalweg profile slope. The response of the Mississippi River to activity of the Monroe Uplift differs slightly from that of the smaller alluvial streams because the Mississippi River channel has been modified by its contact with the uptilted and exposed Tertiary formations. Entrenchment into the Cockfield Sands has occurred within the uplift zone and the channel sinuosity has been modified by the exposed resistant Yazoo clay.

In an actively uplifting area, the channel slope of an alluvial stream does not appear to remain constant but is shown to increase substantially downstream of a domal axis. Therefore it cannot be assumed that a channel is able to maintain a constant or equilibrium channel slope in an area where active tilting of the stream valley is occurring. The Mississippi River has a highly irregular thalweg profile through the Monroe Uplift (Winkley, 1980). The thalweg slope is actually reversed through much of the uplift zone. At river mile 485 (A.H.P.), the mean thalweg slope increases downstream from -0.20 ft/mi.

to +0.52 ft/mi.

In addition to changes in the channel morphology in response to domal uplift or tilting of the valley, streams may also change their entire courses or positions in the valley in response to earth movements. The westward shifting of old Arkansas River channels through time suggests that stream valleys could be displaced through time in the downtilt direction of an actively tilting valley.

The data developed in this section show that the shape, location, sediment load, and drainage basin characteristics of Mississippi River tributaries could be responding to neotectonic activity. In addition, these data provide evidence that lateral shift of a major river, the relic Arkansas River, could be caused by subtle surface movements caused by neotectonic activity.

LOWER VALLEY - WIGGINS UPLIFT

Holdahl and Morrison (1974) and Brown and Oliver (1976) conclude from precise leveling data that subsidence is occurring in the Gulf coastal areas. Rates of subsidence of 4 to 6 mm per year are apparent. In addition Holdahl and Morrison (1974) indicate areas of anomalous subsidence in the New Orleans - Baton Rouge vicinity and at Monroe, Louisiana. These two areas are associated with urban and industrial withdrawal of ground water.

Brown and Oliver (1976) compare the rates of uplift and subsidence evident from geologic data with similar rates evident from precise leveling data, and conclude that present-day movements are up to 2 orders of magnitude larger than average rates determined from long-term geologic record. Thus, it is concluded that movements are either episodic or oscillatory about a long-term trend. This phenomenon of episodic tectonic movement had been recognized earlier in a study of tectonic subsidence of the Mississippi River deltaic plain: "Most movement

probably occurs in spasms, and average rates of movement, which would allow a prediction of the tectonic portion of total subsidence would be very difficult to establish." (U. S. Army, Corps of Engineers, 1958).

TILT RATES

Adams (1980) used leveling to determine tilt rates for the central United States and to demonstrate a strong correlation between tilt rates and adjacent river channel sinuosity. Additional data have been compiled which indicate that the correlation between tilt rate and river sinuosity is valid for the Mississippi River down to Baton Rouge, Louisiana. Below Baton Rouge, river mile 228, there is little correlation between sinuosity and tilt rate (Adams, 1981). The following tabulation shows the correlation of river sinuosity and tilt rate:

RIVER	LOCATION	CORRELATION COEFFICIENT
Mississippi	St. Louis-Cairo	0.95
Mississippi	Cairo-South of Memphis	0.85
Mississippi	Vicksburg-Torras	0.75
Mississippi	Torras-Union	0.72

No tilt rate could be established in the Memphis to Vicksburg reach because precise re-leveling data by the National Geodetic Survey do not exist for this reach. It is strongly recommended that these data be acquired in future surveys. This is the only gap in a first-order level loop along the Mississippi River between Cairo and New Orleans.

ANALYSIS OF NEOTECTONIC EFFECTS

The tilt rates discussed in the previous section were determined from surveys immediately adjacent to the river. In further analysis of possible tectonic-related mechanisms acting in the Lower Mississippi Valley, additional NGS leveling routes were considered.

Figure 5-28 is a graph of bench mark movement versus distance along a survey line from Jackson, Mississippi to New Orleans, Louisiana. The terminuses of the survey line are shown in a map, Figure 5.29. The sharp peak near McComb, Mississippi should be neglected; however, the NGS data indicate a broad uplift in which activity is presently continuing, the magnitude of which is about 3.3mm/yr. maximum for the period 1934 to 1969. The geological literature (Fisk, 1944; Murray, 1961) has reported uplift in the area. Three distinct features have been identified in the vicinity: the Wiggins Uplift, the Southwest Mississippi Anticline, and the Adams County Uplift. The maximum uplift location is near the Mississippi-Louisiana boundary. Fisk (1938) provides additional substantiation of the regional tilting of the area by pointing out that the older (Bentley) terrace is steeper than the younger (Montgomery) terrace along Bayou Sara (for location see Figure 5.31).

The significance of this regional uplift on the Mississippi River can be illustrated by investigation of the change in sinuosity along the river from Vicksburg, Mississippi to Union, Louisiana. The following tabulation lists sinuosities measured from Fisk (1944) maps for the period 1944, 1820, and stage 16, about 1600. The sign (+ or -) between

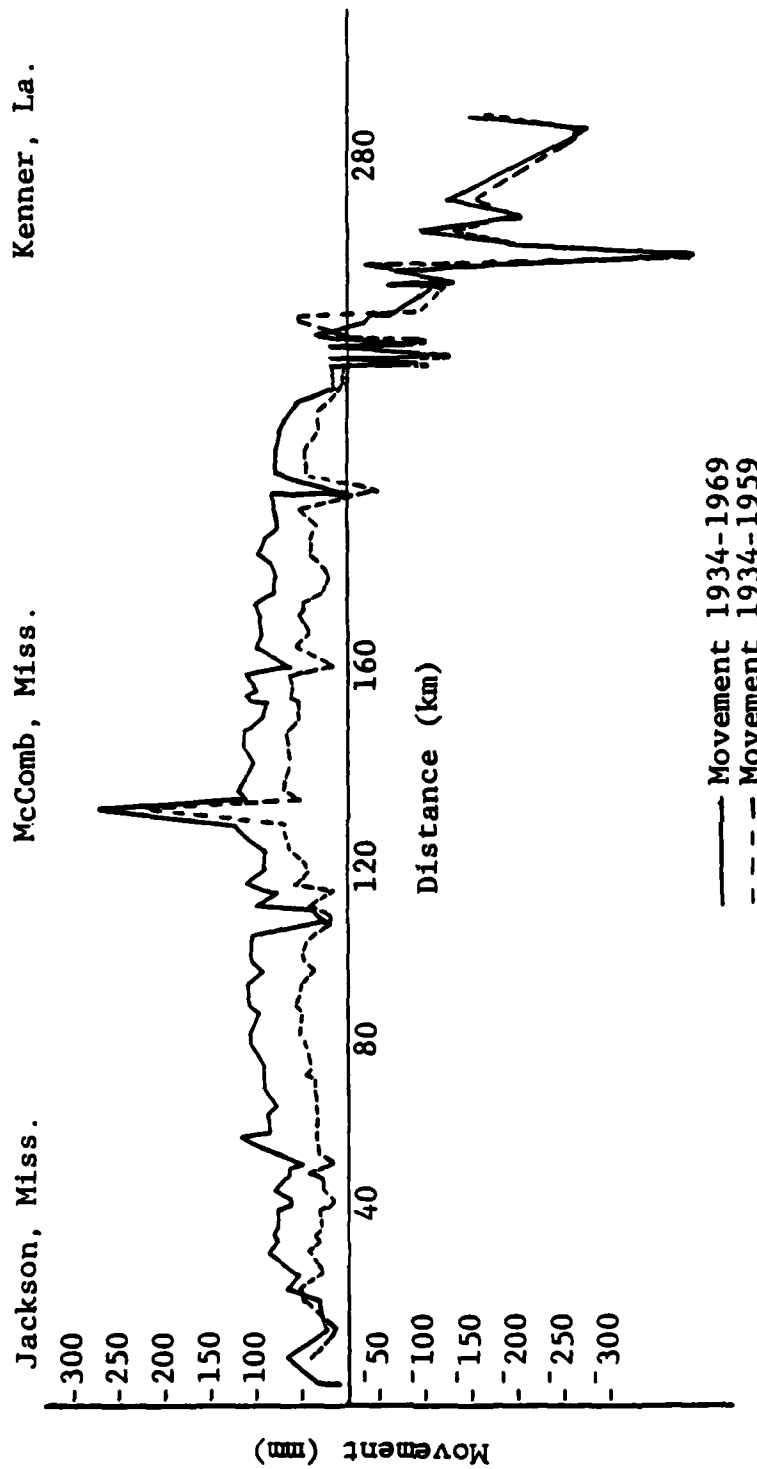


Figure 5.28 Benchmark Movement, Jackson, Miss. to New Orleans, La.

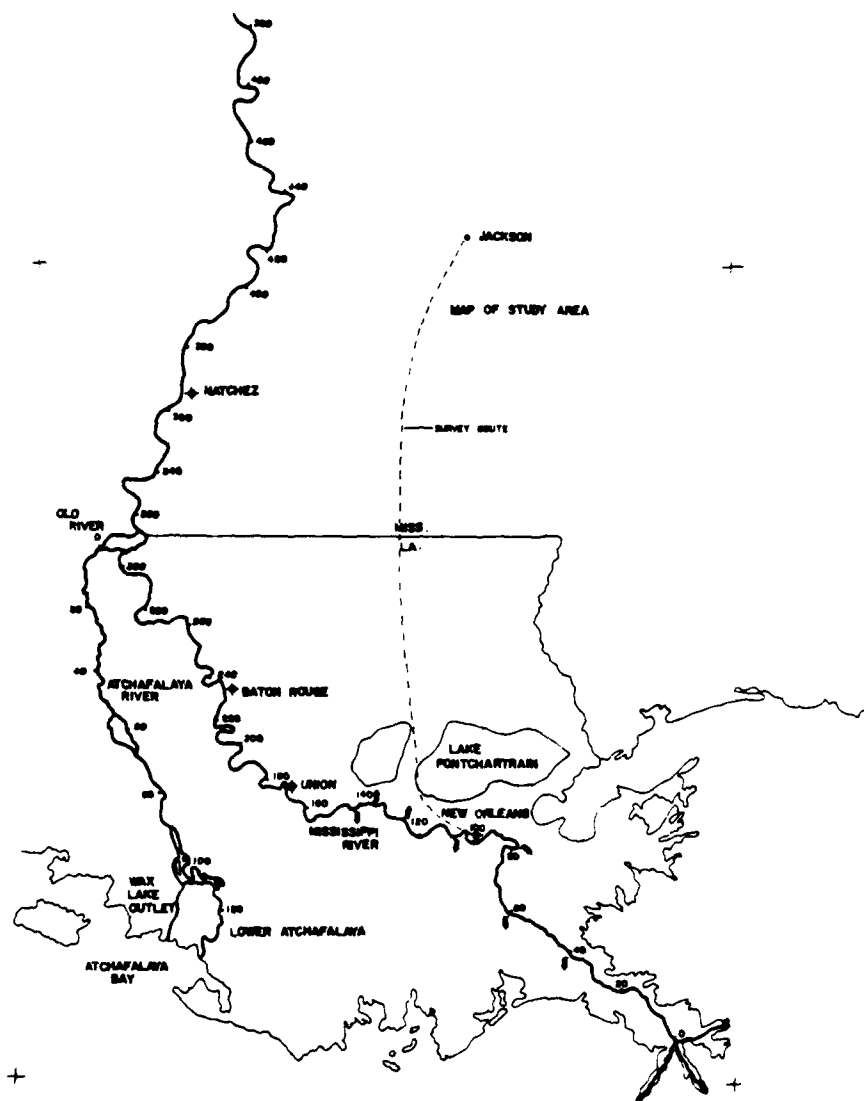


Figure 5.29

values indicates the direction of natural sinuosity change through time.

The asterisk (*) indicates that man-made channel change occurred.

LOCATION	SINUOSITY:				
	<u>1944</u>		<u>1820</u>		<u>1600</u>
Vicksburg to Griffen (R.M. 437 to 430)	1.27	(*)	1.08	-	1.15
Griffen to Hardscrabble (R.M. 430 to 400)	1.30	(*)	2.19	-	2.39
Hardscrabble to Vidalia (R.M. 400 to 363)	1.11	(*)	1.71	-	2.10
Vidalia to Torras (R.M. 363 to 302)	1.39	(*)	1.75	-	1.72
Torras to St. Francisville (R.M. 302 to 265)	1.62	(*)	2.64	+	2.00
St. Francisville to Baton Rouge (R.M. 265 to 228)	1.39	+	1.28	-	2.02
Baton Rouge to Plaquemine (R.M. 228 to 208)	2.10	+	1.80	+	1.67
Plaquemine to Union (R.M. 208 to 167)	1.85	+	1.74	+	1.56

River sinuosity response to the regional uplift indicates that the sinuosity decreased from 1600 to 1820 in the segments upstream of the uplift axis where the valley slope would be decreased by uplift. Little sinuosity change occurred (1.75-1.72) in the Vidalia to Torras segment between 1820 and 1600. The segments below the uplift axis exhibit a tendency to increase sinuosity, the exception being the natural cutoff at False River (R.M. 260) which occurred in 1722.

Examination of the meander patterns (Fisk, 1944) developing up to the time of False River cutoff indicates that the meander was moving northward along the Bluff Hills near Port Hudson, Louisiana. This movement caused the Mississippi River to shift laterally thereby

shortening two tributaries, Thompson Creek and Alligator Bayou which are known to be prolific sand suppliers; the shortening undoubtedly caused incision of the tributary channel. It is likely that the meander was moving in sand deposits and the rejuvenation of the two tributaries likely increased the sediment discharge from the tributaries. The original sand deposits and the ensuing sediment fan heavily influenced the timing and location of this cutoff. This cutoff is also located in the vicinity of the Bancroft fault zone (Murray, 1961), and its proximity suggests a relationship to the cutoff. With the exception of this natural cutoff, the natural trend has been to increase sinuosity which is the expected mechanism to adjust channel slope when given the increased valley slope due to the uplift.

In addition to the effect on Mississippi River sinuosity and the natural trend identified of increasing sinuosity downstream of the uplift and decreasing sinuosity upstream of the uplift, it is equally significant to consider the uplift feature and the regional pattern of surface movement particularly where these movements may have an effect on river behavior. Figure 5.30 shows the apparent iso-vels of surface movement and their relationship to the Mississippi River as determined from NGS precise level surveys. The pattern of surface movement is likely due to a combination of the Wiggins uplift feature, and normal faulting and subsidence.

The western nose of the Wiggins related uplift feature coincides with the Pleistocene expression of the Mississippi structural trough as identified by Jones, et al. (1956) and tends to shift the Mississippi River westward. Also, NGS releveled survey data indicate that the apparent tilting rate along the Atchafalaya River between river mile

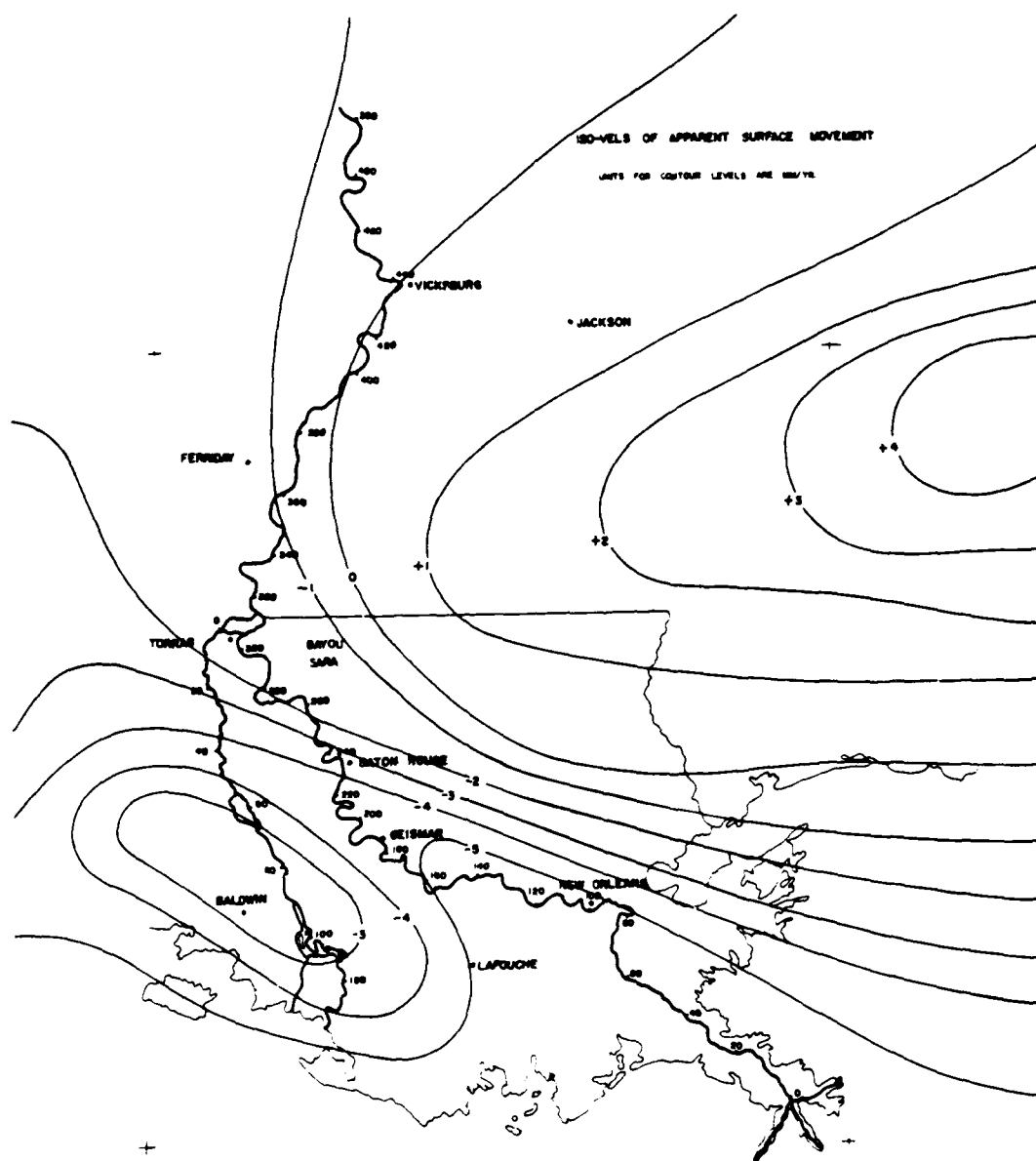


Figure 5.30 (After Holdahl & Morrison, 1974)

zero and river mile 40 is greater than the tilt rate for equal latitudes along the Mississippi River, thus increasing the need for the river to shift to a shorter, steeper course as generally accepted.

Further study of the NGS data to establish relationships of local subsidence in extreme southern Louisiana to regional neotectonics is recommended.

LOCALIZED TECTONIC FEATURES

In addition to the regional uplift and related effect previously discussed, the effects of several localized tectonic features should be considered. The tectonic features shown on Figure 5.31 indicate locations of faults, salt domes, and intrusive rocks. The map was developed from Murray (1961), and from maps published by the Gulf Coast Association of Geological Societies and the American Association of Petroleum Geologists.

A line of five salt domes beginning near the mouth of the Atchafalaya River and extending to the northwest has been very influential in forming previous courses of the Mississippi River. These salt domes first produced a ridge of Pleistocene deposits which, in turn, formed a barrier to the Teche alignment and other courses. Fisk (1944) points out that the five salt domes materially affected the Bayou Teche Route of the Mississippi River. Since the formation of the Bayou Teche Delta Complex, ending about 4000 years before present (Frazier, 1967), no other Mississippi River channel has crossed the Bayou Teche route. The present Atchafalaya River and Wax Lake Outlet are the exception to this.

Of particular interest are the four faults crossing the Mississippi

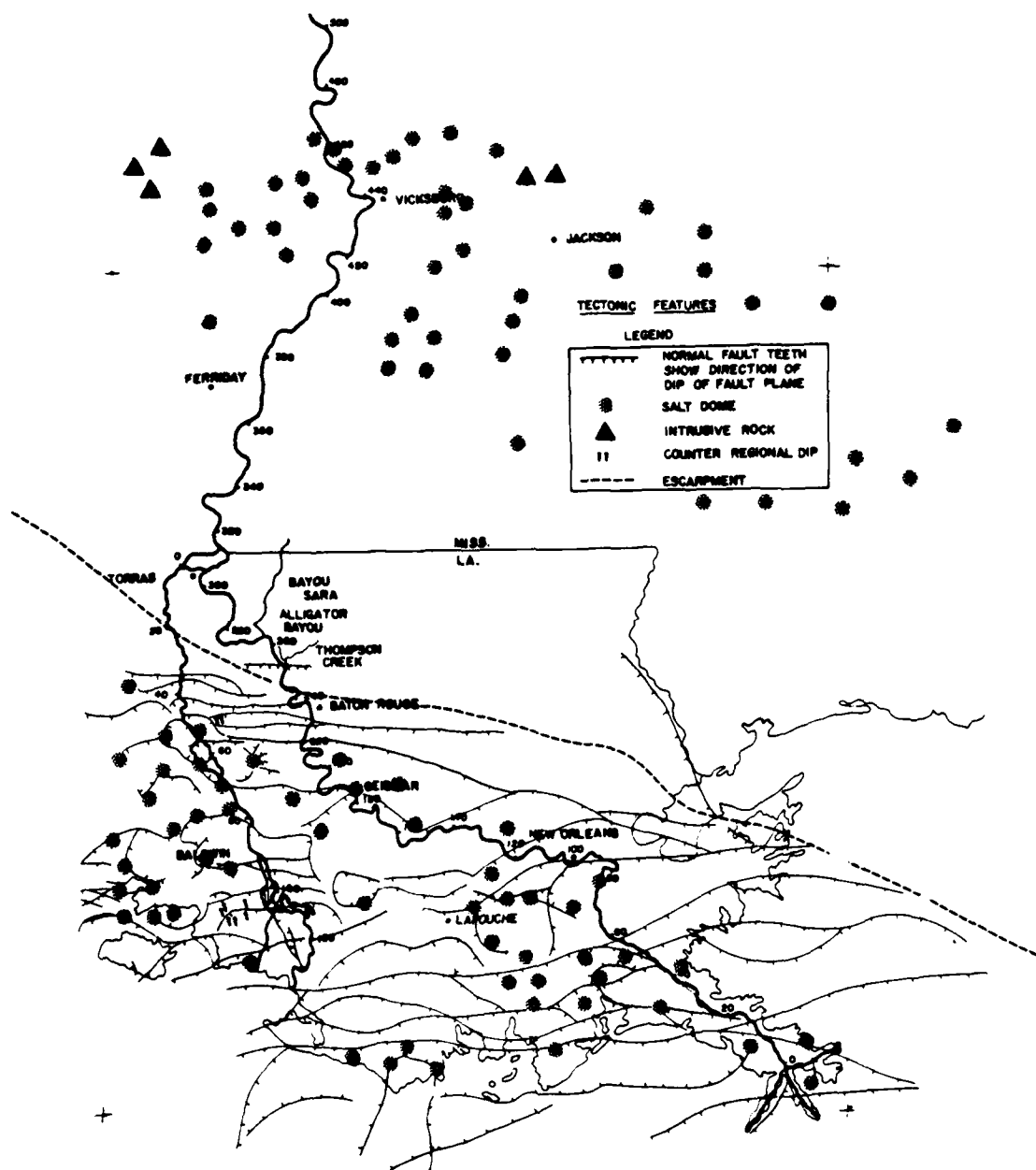


Figure 5.31

River between river mile 220 and river mile 240. As shown on Figure 5.31, the small arrow symbols between two of these faults indicate a counter-regional dip for that fault block. This would have the effect of rotating the block in a direction against the valley gradient. NGS data for river reach from mile 220 to mile 240 indicate that rather rapid movement has occurred across these faults, and tend to substantiate the counter-regional dip in a portion of this system. Further downstream on the Mississippi, from river mile 140 to river mile 180, channel alignment closely parallels a fault zone. And again from river mile 20 to river mile 40 channel alignment closely parallels another fault zone. The two locations imply a strong relationship between river alignment and local tectonic features.

Figure 5.32 is a graph showing the velocity of bench marks along the Mississippi River. The bench mark locations are plotted along a north-south axis with the northern point at a latitude equal to that of Torras, Louisiana. The graph demonstrates a general trend of subsidence with consistent negative bench mark velocity. The graph also indicates a marked tendency between river mile 220 and 240 to exhibit the counter-regional tilting of the surface, i.e., the bench mark velocity trend reverses. The NGS data are for only a short time period and these findings should be viewed as preliminary. Further study of this series of faults, the relationship to groundwater withdrawal in the Baton Rouge vicinity, and the possible episodic nature of fault movement is strongly recommended.

CONCLUSION

This study of neotectonic activity in the vicinity of the Wiggins Uplift has documented the apparent relationship between tilting activity

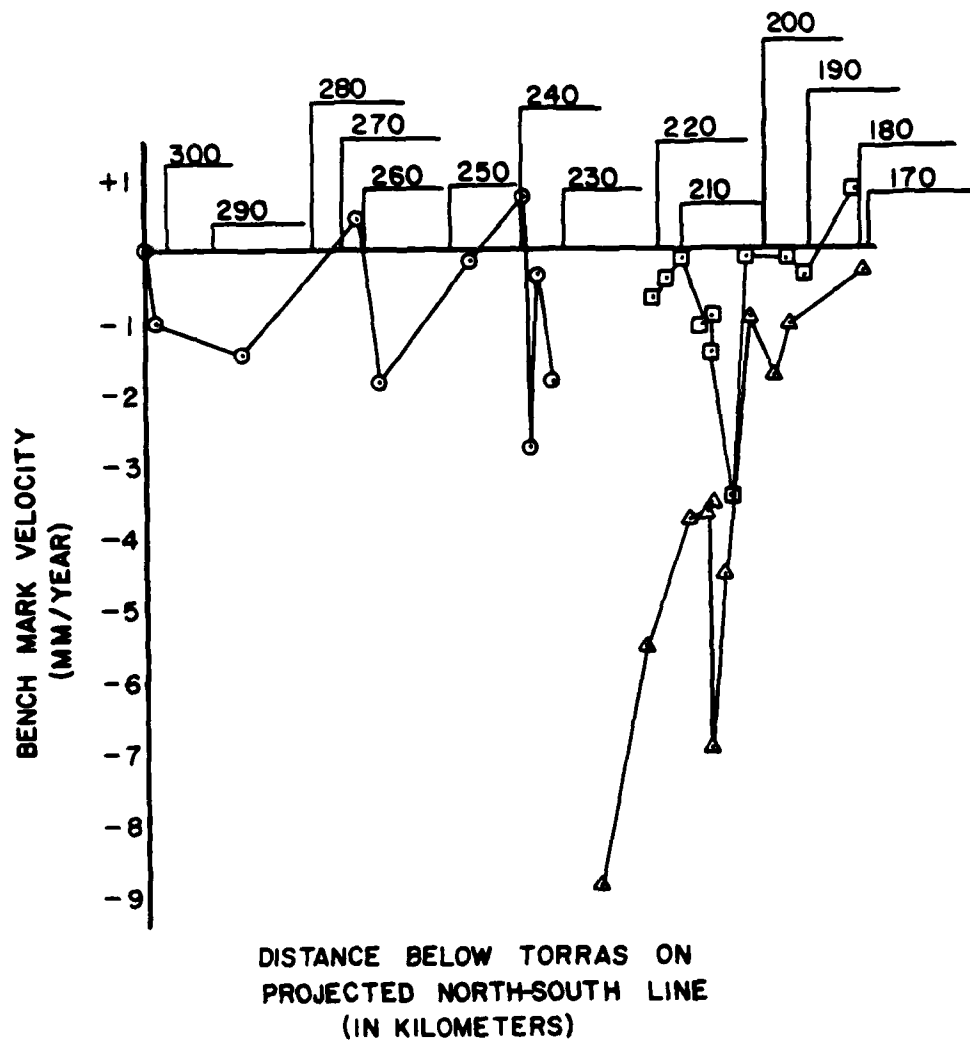


Figure 532

and changing river sinuosity. Obviously the relationships that have been discussed are long term relationships; however, an understanding of these tectonic activities is believed to be important from the standpoint of long term channel stability, channel capacity, and operation of major flood control structures. Additional investigations are needed to better define major relationships and establish reasonable estimates of the magnitude of movements and it is recommended that these investigations be initiated.

CHAPTER 6

THE SIGNIFICANCE OF NEOTECTONICS IN HYDRAULIC ENGINEERING PROJECTS

The preceding chapters have reviewed the geologic literature pertaining to neotectonics and have discussed examples of the effects of movement of various geologic anomalies on the watershed system. In addition, three specific geologic features which are shown to be presently active appear to have an impact on the Mississippi River or tributaries.

Figure 6.1 illustrates the relationship between historic river sinuosity and distance down the Mississippi River Valley. This figure provides a graphic summary of the significance of neotectonics to river engineering projects. Particular attention is directed to the extremes of channel sinuosity south of Cairo, Illinois which coincides with the Lake County Uplift, south of Greenville, Mississippi which coincides with the Monroe Uplift, and north of Baton Rouge, Louisiana which coincides with the Wiggins Uplift. Each of the three uplift features were discussed in the previous chapter.

Investigations indicate that reaches of the river affected by uplift features can experience aggradation upstream of the uplift axis and degradation downstream of the uplift axis. These processes can affect the success of channel improvement plans and also affect long term flood conveyance of the channel.

In general, the effect of neotectonic movement in hydraulic engineering projects can be summarized as follows:

1930-1932
 1861-1893
 1820-1830
 1764
 RIVER CHANNELS

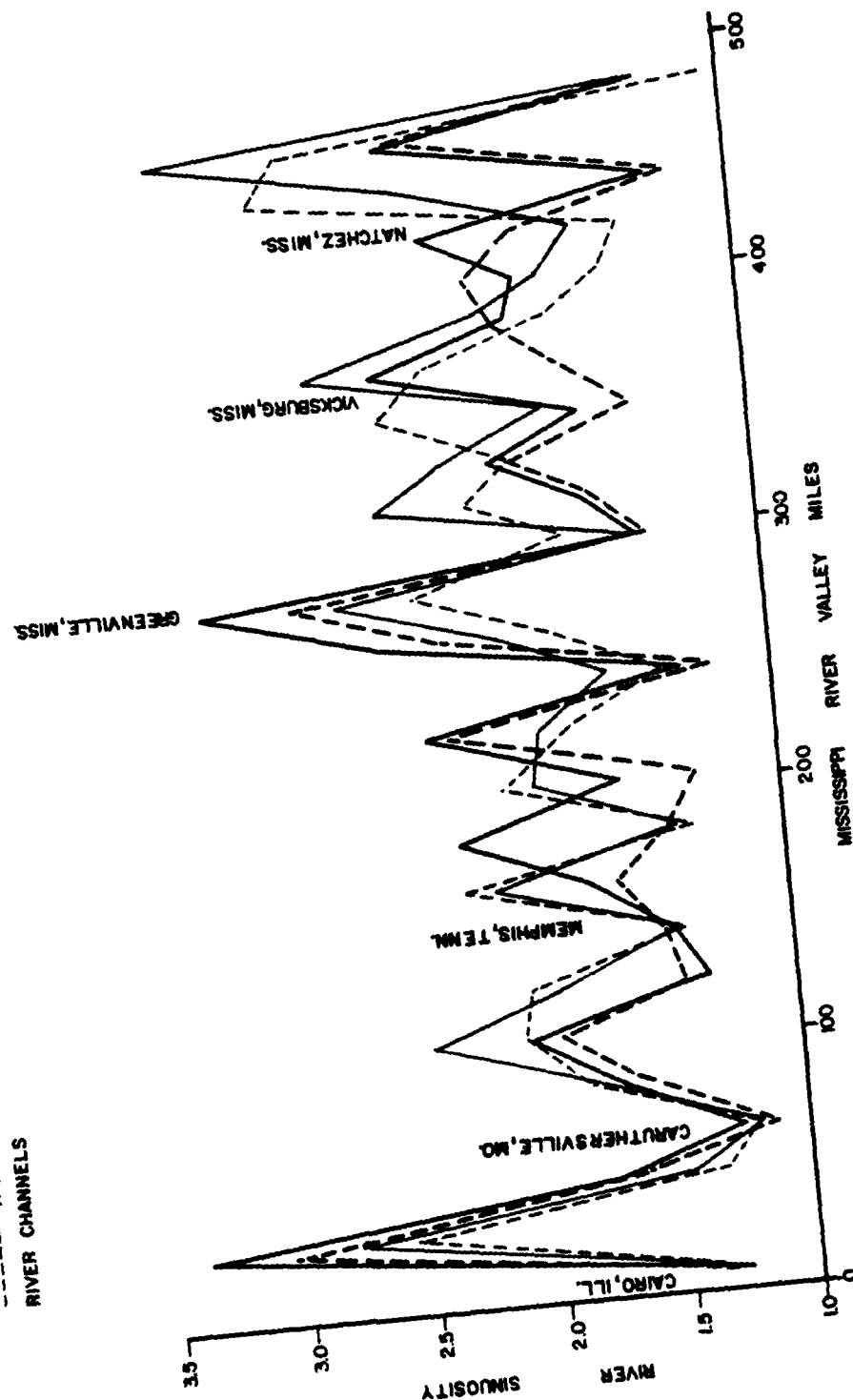


Figure 6.1

- 1.) Change in watershed drainage pattern
- 2.) Channel aggradation or degradation
- 3.) Change in channel pattern or sinuosity
- 4.) Channel diversion or avulsion
- 5.) Flooding due to subsidence.

Examples of each of these effects can be found within the limits of the Lower Mississippi Valley Division. A thorough discussion of these effects has been presented in chapters two and three.

The neotectonic effects altering a watershed drainage pattern are shown in Figure 2-4, and were discussed by Russell (1939). The effect demonstrated in Figure 2-4 is that the increased slope in the lower portion of the drainage basin has changed the poorly defined drainage in the upper section to an efficient dendritic drainage pattern.

In addition, the area above the Wiggins Uplift as shown on the Natchez 1:250,000 quadrangle map indicates the deflection of the Homochitto River, the Pearl River, and the Mississippi River from the uplift region. On a smaller scale, it may be assumed that many tributary channels have equal or greater influence from the Wiggins Uplift.

The significance of altering the drainage pattern as shown in Figure 2-4, or as may be found in further investigation of the Pearl River and Amite River basins, is that decreased hydraulic efficiency upstream of the uplift and increased hydraulic efficiency downstream of the uplift can result. The conditions can result in poor drainage for agricultural land or urban developments in the upper watershed. Likewise, the lower portion of the basin can be subject to increased sediment load and increased stream instability due to increased stream gradient.

The effects of channel aggradation or degradation should be very clear to the engineer familiar with response to alluvial channels. Figure 3-4 shows the relationship between stream response and tectonics. As the slope is increased on the downstream side of a relative uplift, stream power increases and the competence of the channel to degrade is increased. Aggradation results upstream due to reduced gradient, and can result downstream due to oversupply from the degrading reach. Examples of this effect are cited in the Monroe Uplift Region discussion in Chapter Five.

Figure 6.2 provides another way to illustrate the significance of neotectonics in hydraulic engineering projects. This figure was compiled from Mississippi River data provided by the Potamology Section, Vicksburg District, U.S. Army Corps of Engineers and has been verified in experimental investigation (Schumm and Khan, 1972). Scatter of data points about the curve reflect sinuosity variation due to meander growth and natural cutoffs. Depending upon the existing location of a particular river reach on the curve shown in Figure 6.2, a change in valley slope of 0.2 feet (60 mm) could significantly change channel pattern and sinuosity. For example, 3 mm/year uplift for a twenty year period could result in a river pattern change from meandering to braided.

In a closely aligned river like the Mississippi, the river pattern may be fixed by dikes and revetments, but the tendency of a braided channel to form mid-channel bars would be increased by the uplift. The river response to the uplift could be a very troublesome navigation reach requiring high dredging frequency.

Two examples of stream diversion or avulsion are alluded to in this

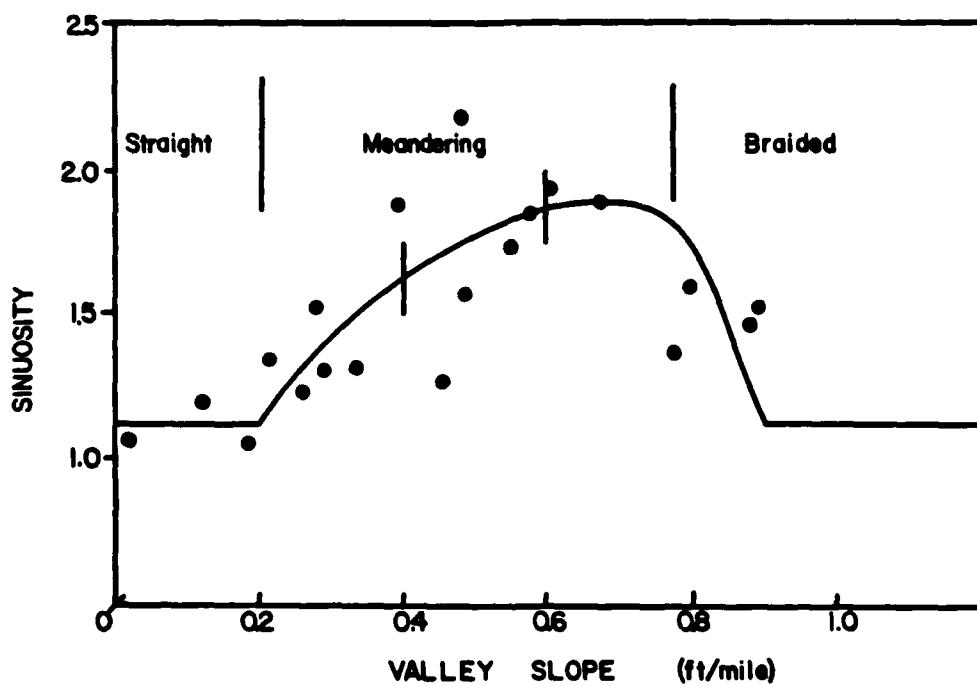


Figure 6.2 Relationship between Valley Slope and Sinuosity, Mississippi River.

report. Geomorphologic evidence indicates that the Monroe Uplift could have been a primary cause in the diversions of the Arkansas River to cause the sequence of channels now present in Northeastern Louisiana. Figure 5-22 shows the routes of relic Arkansas River channels as it shifted in a westward direction in response to continual uplift and shifting of the Monroe Uplift. Continued uplift may have forced the Arkansas River against the western upland and finally forced abandonment of the southerly route at Pine Bluff to the present course. An equally dramatic influence may now be occurring on the Mississippi River from the Wiggins Uplift.

Although coastal subsidence can be attributed to several causes, tectonics is an influence. More important is the fact that the effects and the methods of investigation to determine movement are the same as for upland locations. NGS survey data document subsidence rates on the order of 5-6 mm/year in the Mississippi Deltaic plain. At this rate a total subsidence of about one foot could occur in a 50 year project life.

This final chapter discussing the significance of neotectonics in hydraulic engineering projects is intended as a review of effects more completely discussed in previous chapters of the report. The hydraulic engineer designing projects along the alluvial system has already recognized the fact that the river is not a rigid boundary conduit; the channel erodes laterally and scours and aggrades vertically. This preliminary study of the effects of neotectonics has identified another degree of freedom in the alluvial system, i.e., the earth platform on which the river system flows may also be dynamic.

Continued investigation of neotectonic effects should include

further identification of active areas within the Lower Mississippi Valley Division. Both the Pearl River and the Red River flow over recognized uplift features and their associated fault systems. It is certain that these two rivers, plus the Ouachita River, and portions of the Mississippi River main stem are influenced significantly by neotectonic movement. Planning for navigation and flood control projects should consider of the effects of neotectonic movement.

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